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	Engineering and Design RUNOFF FROM SNOWMELT	
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DEPARTMENT OF THE ARMY
U.S. Army Corps of Engineers
Washington, DC 20314-1000

EM 1110-2-1406

CECW-EH

Manual
No. 1110-2-1406

31 March 1998

Engineering and Design
RUNOFF FROM SNOWMELT

1. Purpose. The purpose of this manual is to provide guidance for computing basin snowmelt runoff as is necessary in the design and operation of USACE water control projects.

2. Applicability. This manual applies to all USACE Commands having responsibility for design of civil works projects.

3. Distribution Statement. Approved for public release, distribution is unlimited.

FOR THE COMMANDER:

ALBERT J. GENETTI, JR.
Major General, USA
Chief of Staff

This manual supersedes EM 1110-2-1406, dated 5 January 1960.

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Chapter 1
Introduction

1-1. Purpose

This manual provides technical background and guidance for computing basin snowmelt runoff as is necessary in the design and operation of U.S. Army Corps of Engineers (USACE) water control projects. This manual discusses the basic theoretical principles of snow hydrology and the practical applications of this theory in forecasting and design. It summarizes several important snowmelt runoff models and offers guidelines for model selection. This manual represents an update of EM 1110-2-1406, Runoff from Snowmelt, dated 5 January 1960, which is now obsolete. While many of the basic principles and techniques presented in that manual have been retained, numerous advancements in computer, communications, and data acquisition technologies are now reflected. This manual is applicable to USACE offices in which snow hydrology considerations affect runoff and streamflow derivations.

1-2. Background

In the mid-1940s, the Federal Government initiated a major research program as a cooperative effort between the U.S. Army Corps of Engineers and U.S. Weather Bureau, with the major impetus being to develop procedures to derive spillway design floods for the major dams that were being planned for western river basins subject to snow runoff. The *Cooperative Snow Investigation Program* established three snow laboratories that were operated until the mid-1950s. The Central Sierra Snow Laboratory was located in the Sierra Mountains of California near Donner Pass; the Upper Columbia Snow Laboratory was located in Glacier National Park in Montana; and the Willamette Basin Snow Laboratory was in the upper McKenzie River drainage in Western Oregon. (The Central Sierra Laboratory continues to be operated by the Department of Agriculture.) The results of the laboratory experiments and other scientific research of the program were documented in numerous Technical Reports, Research Notes, and Technical Bulletins. These were in turn compiled into a summary report, *Snow Hydrology* (U.S. Army Corps of Engineers 1956). This document remains a valuable resource for hydrologists and engineers working with snow hydrology applications. The final product of the *Cooperative Snow Investigations Program* was EM 1110-2-1406, the predecessor of this document. Since the 1960s, advances in applied snow hydrology have centered primarily around computer applications of the methodologies developed by USACE and subsequent researchers. These include the following:

a. Development of many conceptual snowmelt models.

b. Use of new technology to acquire data for measuring various aspects of snow.

c. Employment of computers in managing and analyzing hydrometeorological data.

d. Use of new communications technologies for rapid access to data, even in the near real-time.

With all these changes, snowmelt models are now internalized in operational forecasting more than ever before, and their future use will increase as more efficient capabilities for data acquisition, communications, and analysis are developed.

1-3. Snow Hydrology Modeling

In this manual, focus is placed primarily on computing runoff and streamflow in which snow has played a role in the process. This computation, typically accomplished with a computer model of some sort, includes the following considerations.

a. Collection and handling of competent spatial and temporal data for model input. This operation, critical especially in real-time forecasting, has been enhanced in recent years by the development of remote sensing and the availability of near-real-time water control data.

b. Formulation of the structure of the snowmelt model. How the model deals with the complex physics of accumulation, snowmelt, areal snow distribution, and snow-soil interactions must be defined so that new data collection and handling techniques can be rationally analyzed and incorporated

as input. This step involves selecting a computer program that is appropriate for the application, then using the options available correctly and intelligently.

c. Application of the model in either analysis or forecasting. Here the skill and experience of the user come into play as a model is calibrated, tested, then applied in the intended application. This cannot be done without a thorough background in snow hydrology, making use of basic principles that are described in this manual and in other references.

1-4. References

Related publications include:

a. ER 1110-2-248 Requirements for Water Data Transmission Using GOES/DCS

b. ER 1110-2-249 Management of Water Control Data Systems

c. EM 1110-2-1415 Hydrologic Frequency Analysis

d. EM 1110-2-1416 River Hydraulics

e. EM 1110-2-1417 Flood-Runoff Analysis

f. EM 1110-2-3600 Management of Water Control Systems

1-5. Bibliography and Definitions

A bibliography of other reports and important papers pertaining to snowmelt runoff that are cited in the text is provided in Appendix A of this manual. Additionally, a glossary of terms and definitions is included in this Engineer Manual as Appendix B. A comprehensive listing of literature pertaining to snow is contained in the *Bibliography on Cold Regions Science and Technology* that is periodically published by the U.S. Army Cold Regions Research and Engineering Laboratory and the Library of Congress. By regularly reviewing this Bibliography, the user can efficiently keep abreast of continuing developments in the field of snow hydrology to supplement the contents of this manual.

1-6. Scope and Content

This manual includes both theoretical and practical topics. The basic theoretical concepts of snow hydrology are presented in Chapter 2, "Snowmelt Runoff—A Review of the Fundamental Processes." This chapter draws upon *Snow Hydrology* and more recent research work to summarize the physical processes involved in snow accumulation, metamorphosis, and melt, and to present fundamental equations that describe these processes. After a discussion of data collection and analysis in Chapter 3, the physical processes are again discussed, this time with regard to practical applications in forecasting and design, in Chapters 4 through 9. Chapter 4, "Snow Accumulation and Distribution," discusses techniques—both simple and complex—that can be used to estimate snow quantity and areal extent at the beginning of a snowmelt runoff event. Chapter 5, "Snowmelt—Energy Budget Solutions," presents the semiempirical equations that have been developed from the basic theoretical principles for use primarily in the derivation of design floods in a snow environment. In Chapter 6, "Snowmelt— Temperature Index Solutions," the simpler alternative method of estimating snowmelt rates, used widely in real-time hydrologic forecasting, is discussed. Chapter 7, "Effect of Snow Condition on Runoff," covers the practical considerations associated with the metamorphosis of snow—how the condition of the snow can affect the determination of runoff. Chapter 8, "Snowmelt—Accounting for Changes in Snow and Snowcover," describes approaches to modeling the change of snow quantity and areal extent during snowmelt. Chapter 9, "Statistical Analyses," summarizes statistical techniques that are commonly used in snow hydrology.

a. The techniques and "tools" described in Chapters 4 through 9 are further described in terms of their use in practical engineering applications in Chapter 10, "Snowmelt Runoff Analysis for Engineering and Forecasting Applications." Examples include simple and complex derivations of design floods, reservoir operational analysis, and operational forecasting. In Chapter 11, "Guidelines for Snowmelt Model Selection," available operational models are described.

b. In addition to Appendixes A and B noted above, several other appendixes provide detailed technical material to augment the information presented in the main body of the manual. Appendix C, "Summary of Basic Physics Principles—Heat, Heat Transfer, and Thermal Properties of Water," summarizes some basic physics of water that are applicable in snow hydrology. Included are some basic tables of physical properties in both English and SI units. Appendix D, "Meteorological Relationships," presents a number of charts drawn from the *Cooperative Snow Investigation Studies* that are useful in describing the influence of meteorological phenomena on snowmelt. Appendix E presents SI unit versions of the generalized energy budget equations that are discussed in Chapter 5. Finally, Appendix F, "Summary Descriptions of Selected Operational Snowmelt Models," summarizes the characteristics of several widely used computer models that can be used to simulate snowmelt runoff.

1-7. Units

The equations in this manual will be presented in both SI and English units. If the reader refers to modern textbooks on physics and meteorology, the SI convention would be used exclusively. However, once the discussion involves the experimental relationships that were developed in the 1950s, a shift to current U.S. practice (English units) must be made. Further discussion of units can be found in Paragraph 5-2*b*.

Chapter 2
Snowmelt Runoff—A Review of Fundamental Processes

2-1. General Introduction

In many regions of the United Sates, snowfall and the resulting seasonal snowcover represent an important source of water. When the snowpacks melt, the snowmelt recharges the groundwater and replenishes surface water storage. Excessive snowmelt runoff can cause flooding, while inadequate snowmelt is often the prelude to later drought.

a. When snow melts, the ice that composes the snow is converted into water. This water is called snowmelt. Since the conversion from ice to water requires the input of energy (or heat), the process of snowmelt is inextricably linked to the flow and storage of energy into and through the snowpack. These linkages between the flow and storage of both water (i.e., ice and liquid water) and energy (or heat) are summarized in Figure 2-1 to facilitate the discussion and to clarify the complicated processes that control snowmelt runoff.

b. The sources of energy that cause snowmelt include both shortwave and long-wave net radiation, convection from the air (sensible energy), vapor condensation (latent energy), and conduction from the ground, as well as the energy contained in rainfall. These energy fluxes are shown in the upper left of Figure 2-1 and are labeled Q_{sn}, Q_{ln}, Q_h, Q_e, Q_g, and Q_p, respectively. These fluxes are usually measured as energy per time per unit area of snow. The energy budget equation that describes the energy available for snowmelt is given in Equation 2-1 below. The total energy available for snowmelt is Q_m.

$$Q_m = Q_{sn} + Q_{ln} + Q_h + Q_e + Q_g + Q_p - \Delta Q_i \qquad (2\text{-}1)$$

ΔQ_i is the rate of change in the internal energy stored in the snow per unit area of snowpack. This term is composed of the energy to melt the ice portion of the snowpack, freeze the liquid water in the snow, and change the temperature of the snow. Thus, during periods of warming, the net flux of heat (ΔQ_i) is into the snow, while during periods of cooling, the net flux (ΔQ_i) is out of the snowpack. Therefore, the amount of energy available to cause snowmelt varies and can be dynamic, depending on the magnitudes of the various energy inputs to the snowpack. Male and Gray (1981) suggest that snowmelt is not homogeneous throughout the snowpack depth and point out that most of the melting occurs at the upper and lower interfaces of the snow (i.e., the interfaces with the atmosphere and the ground).

c. Whenever sufficient energy is available, some snow (ice) will melt and form liquid water (i.e., snowmelt). Since the physical structure of the snowpack is a porous matrix, this snowmelt will be held as liquid water (provided it does not refreeze) in the interstices between the snow grains and will increase snow density and snow water content. The snowpack is commonly called "ripe" when it is isothermal at 0 °C and saturated. Whenever the capacity of the snowpack interstices to hold the liquid water is exceeded, some of the snowmelt will begin to move down-gradient (called direct surface runoff in Figure 2-1) to become a portion of the snowmelt runoff. Additionally, some of the snowmelt may infiltrate into the ground. The amounts that infiltrate depend on inherent soil characteristics, the soil moisture content, as well as whether or not the ground surface is frozen. The infiltrated snowmelt later reemerges as interflow into stream channels, or it percolates into deeper groundwater storage. These snowmelt pathways are delineated in Figure 2-1.

d. Estimates of snowmelt amounts are derived through the use of energy balance equations or by some empirically defined snowmelt index. Determinations of the amounts and the temporal distributions of snowmelt runoff require additional analysis of the storage of the snowmelt in the snowpack and transmission of the snowmelt through the snowpack as well as along the surface of the ground as it courses its way to the stream channel.

e. This chapter will discuss the theoretical basis for snowmelt at a point and from a basin or watershed. Throughout, the overall energy and water mass pathways shown in Figure 2-1 will form the framework for the discussion.

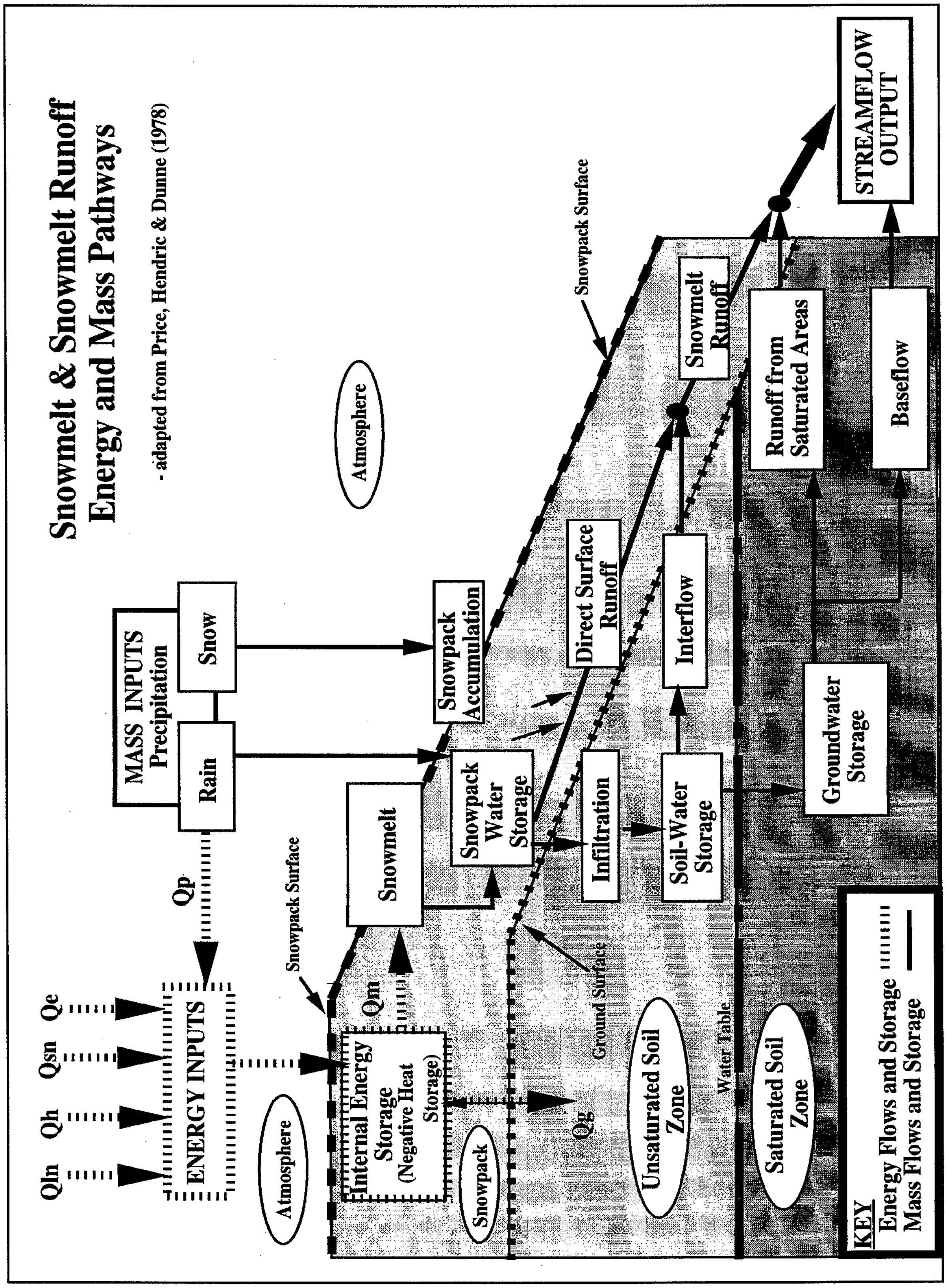

Figure 2-1. Schematic of the snowmelt process. (After Price, Hendrie, and Dunne (1979))

2-2. Energy and Mass Balance of the Snowcover

Evaluating snowmelt theoretically is a problem of heat transfer involving radiation, convection, condensation, and conduction. The relative importance of each of these heat transfer processes is highly variable, depending upon conditions of weather and the local environment. Gray and Prowse (1992) tabulate selected results of the relative contributions of each heat transfer process as a function of site environment. The basic equations and coefficients that describe snowmelt at a point have been derived primarily from various laboratory and field experiments.

a. General. Equation 2-1 summarized the energy sources available to melt snow. The summation of all sources of energy (heat) represents the total amount of energy available for melting the snowpack (Q_m). The amount of snowmelt at a point may be expressed by the general formula given as Equation 2-2

$$M = \frac{Q_m}{334.9 \rho_w B} \tag{2-2}$$

where

M = snowmelt, mm of water equivalent

Q_m = algebraic sum of all heat components, kJ/m2

B = thermal quality of the snow (e.g., ratio of heat required to melt a unit weight of the snow to that of ice at 0 °C)

334.9 = latent heat of fusion of ice, kJ/kg

ρ_w = density of water, kg/m^3

(1) Equation 2-2 may also describe the snowmelt per unit time (for example, mm water equivalent day) when Q_m is expressed in kJ/m^2 per day.

(2) A melting snowpack consists of a mixture of snow (ice) and a small quantity of free (liquid) water trapped in the interstices between the snow grains. The relative proportion of a snowpack that consists of ice determines the thermal quality (B) of the snowpack. A snowpack that contains no free water has a thermal quality of 1.0. However, after melt has begun, there is some free water held within the snow matrix, yielding a thermal quality of less than 1.0. The heat energy required to release 1 g of water is somewhat less than the latent heat of fusion of water (that is the energy required to change state from ice to water; 334.9 kJ/kg or 80 cal/g for pure ice). For a melting snowpack, after free drainage by gravity for several hours, the thermal quality normally averages between 0.95 and 0.97, corresponding to a 3- to 5-percent liquid water in the snow.

(3) The thermal quality of snow may be far lower for "ripe" snows and in extreme cases where the water cannot drain freely from the snowpack.

b. Radiational energy exchange. Radiational energy is the prime source of energy at the Earth's surface. Some of this energy is classed as solar or shortwave radiation (radiation having wavelengths (λ) between 0.2 and 2.2 μm) and terrestrial or long-wave radiation (wavelengths between 6.8 and 100 μm). The first two terms of Equation 2-1 are sometimes referred to as net radiation Q_n, the sum of net shortwave Q_{sn} and net long-wave Q_{ln} energy fluxes. As the net long-wave exchange is often a loss from the snow surface, Q_n is expressed as

$$Q_n = Q_{sn} - Q_{ln} \tag{2-3}$$

(1) Shortwave radiation is the most important source of energy to the snowpack. The net amount of radiant energy that is available to melt snow depends on how much of the radiation is either reflected from or absorbed by the snowpack. The amount of heat transferred to the snowpack by solar radiation varies with latitude, season, time of day, atmospheric conditions, forest cover, and reflectivity of the snow (albedo). The intensity of incident solar radiation just above the Earth's atmosphere and normal to the path of radiation is virtually constant at 1.35 kJ/m^2 per second, the solar constant. In general, less than 50 percent of this incident solar radiation reaches the Earth's surface. As solar radiation passes through the Earth's

atmosphere, it is attenuated through reflection off clouds, scattered by air molecules and particulates in the atmosphere, and absorbed by a number of molecular structures contained in the atmosphere. By far, the greatest change in the portion of solar radiation transmitted through the atmosphere is caused by varying cloud cover, so that direct measures of solar radiation at the ground surface principally show the effect of depletion by clouds. Inasmuch as such measurements are not generally available at a given location, it usually becomes necessary to estimate incoming radiation indirectly from duration of sunshine data, observations of cloud conditions, or diurnal air temperature fluctuations. See Appendix D for further discussion of radiational energy exchange, including several charts showing how radiational flux varies.

(2) Additionally, the local environment has a marked effect upon the amount of solar radiation received on the snow surface. The relative ratio of the daily solar radiation incident upon a snow surface to that on a horizontal surface is a function of the surface slope angle to north or south (or aspect), the latitude, the season, and the amount of diffuse sky radiation relative to direct solar radiation. More complete descriptions of methods for calculating incident radiation and the effects of local environment are given by List (1968), Dozier (1979), Oke (1978), Male and Gray (1981), and Gray and Prowse (1992).

(3) Forest cover can also play an important role in the amount of solar energy that reaches the snow surface. For example, in coniferous forests, the transmission percentage varies with the type, density, and condition of trees. Transmission also varies with the season, because of the change in the shading effect of the trees with the solar altitude. The determination of the amount of sunshine transmitted through the forest is at best an approximation.

(4) The reflectivity of the snow surface plays an important role in the amount of energy available to cause snowmelt. Large portions of the shortwave radiation that reach the snow surface can be reflected. Since snowpack reflectivity varies over a considerable range, it is an important consideration in estimating the amount of solar energy absorbed by the pack. Albedo (A) is defined as the percentage of the incident shortwave radiation that is reflected from the snow surface. Values for albedo range from more than 80 percent for new-fallen snow to as little as 40 percent for melting, late-season, ripe snow.

(5) The amount of energy available for snowmelt from the absorption of shortwave radiation (Q_s) is

$$Q_s = (1 - A) I_i \tag{2-4}$$

where

A = albedo (expressed as a decimal fraction)

I_i = daily incident solar radiation (kJ/m^2 per day)

(6) In the middle latitudes during late spring, the maximum solar radiation for a clear day on a horizontal surface is about 52 MJ/m^2. With a minimum albedo of 40 percent, the resulting possible shortwave radiation melt for an unforested area is on the order of 6.4 cm/day. However, some of the energy absorbed by the snowpack from solar radiation is radiated from the snowpack to the atmosphere as long-wave radiation. Snow is nearly a perfect blackbody, with respect to long-wave (terrestrial) radiation, absorbing all such radiation incident upon it and emitting the maximum possible radiation in accordance with the Stefan-Boltzman law (Equation 2-5).

$$Q_l = \epsilon \sigma T_s^4 \tag{2-5}$$

where

Q_l = total shortwave energy emitted by the snow, kJ/m^2 per second

ϵ = 0.99 for clean snow

σ = Stefan-Boltzman constant, 5.735 × 10^{-11} kJ/m^2 s K^4

T_s^4 = blackbody temperature in Kelvin (K) (temperature of the snow surface)

(7) Consider a melting snowpack having a surface temperature of 0 °C. According to the Stefan-Boltzman law, the snowpack will lose energy at the rate of 0.315 kJ/m^2 per second. Opposed to this is the back-radiation, or long-wave radiation, reflected back from the atmosphere or the forest cover. For clear skies, the heat gain from back-radiation is generally less than the heat loss, so that there is net heat loss from the snowpack by long-wave radiation. With cloudy skies or beneath a forest canopy, however, the back-radiation may be greater or less than the loss from the snowpack, depending principally upon the ambient air temperature.

c. Turbulent transfer.

(1) Energy is also exchanged between the snow-pack and atmosphere through the processes of convection and condensation. Depending on the climatological and local weather conditions, the relative importance of these processes differs widely. For example, during clear weather in the spring, energy exchange by the process of turbulent exchange from the atmosphere is of secondary importance compared with radiation for snowmelt. However, during a winter rain on snow, turbulent exchange is the dominant heat exchange process. Turbulent exchange involves the transfer of sensible heat from warm air advected over the snowfield (convection), and also the latent heat of condensation of water vapor from the atmosphere condensed on the snow surfaces. Computation of the transfer of sensible and latent heat from the atmosphere is complex from a theoretical standpoint, and exchange coefficients are derived empirically from controlled experiments.

(2) The principal variables affecting convective (sensible) heat exchange are the temperature gradient of the atmosphere measured above the snow surface and the corresponding wind speed. Similarly, the primary variables affecting condensation (latent) heat exchange are the vapor pressure of the atmosphere and the snow surface and the corresponding wind speed. Equations 2-6 and 2-7 describe sensible and latent heat transfer, respectively (Gray and Prowse 1992).

$$Q_h = D_h u_z (T_a - T_s) \tag{2-6}$$

$$Q_e = E_e u_z (e_a - e_s) \tag{2-7}$$

where

D_h = bulk transfer coefficient for sensible heat transfer, kJ/m^3 °C

u_z = wind speed at a chosen height above the snow surface, m/s

T_a = temperature at the air surface, °C

T_s = temperature at the snow surface, °C

D_e = bulk transfer coefficient for latent heat transfer, kJ/m^3 Pa

e_a = vapor pressure of the air surface, Pa

e_s = vapor pressure of the snow surface, Pa

d. Heat conduction from the ground. Heat entering the snow from the ground (Q_g) by solid conduction is a very small component to the overall energy budget, especially compared with the radiational and turbulent exchange at the air/snow interface. This ground heat component can be neglected over short periods of time (less than 1 week). Although the daily melt caused by ground heat is small, it can amount to a significant quantity of water over an entire snow season. Most lumped, conceptual models use constant daily values in the range of 0-5 J/m^2 per second. Ground heat flow can also be estimated using soil temperature gradients measured near the surface in an equation for steady-state, one-dimensional heat flow by conduction:

$$Q_g = k \frac{dT_s}{dz} \tag{2-8}$$

where

k = thermal conductivity of the soil

$\frac{dT_g}{dz}$ = temperature gradient from soil to snow

e. Heat convected by rain. The heat convected from the snow by rainfall is

$$Q_p = C_p \rho_w P_r (T_r - T_s)/1000 \qquad (2\text{-}9)$$

where

C_p = specific heat of rain, kJ/kg °C

ρ_w = density of water, kg/m^3

P_r = rain quantity, mm/unit time

T_r = temperature of the rain, °C

T_s = snow temperature, °C

The temperature of the rain is assumed to be equal to the air temperature or, if available, the wet-bulb temperature. The specific heat C_p is equal to 4.20 kJ/(kg °C) for rainfall and 2.09 kJ/(kg °C) for snowfall.

f. Internal energy. By definition, if the cold content or heat deficit of the snowpack is positive, the snowpack's temperature is below freezing. The internal energy Q_i can be changed and the heat deficit reduced by the heat released when melt or rainwater freezes within the snow cover. This phenomenon is prominent during diurnal temperature cycles with refreezing at night because of radiational cooling. Melt and rainwater will continue to freeze within the snow cover until the total heat deficit reaches zero. When the total heat deficit reaches zero, the snow cover will become isothermal at 0 °C. This internal energy is calculated by the following expression (Gray and Prowse 1992):

$$Q_i = d_s (\rho_i C_{pi} + \rho_l C_{pl} + \rho_v C_{pv}) \, T_m \qquad (2\text{-}10)$$

where

d_s = depth of snow

ρ = density, ρ_i = 922 kg/m^3, ρ_l = 1000 kg/m^3

C_p = specific heat, $C_{pi} \cong$ 2.1 kJ/kg °C; $C_{pl} \cong$ 4.2 kJ/kg °C

T_m = mean snow temperature, °C

The subscripts *i*, *l*, and *v* refer to the ice, liquid, and vapor phases, respectively. The contribution of the vapor phase is assumed negligible.

2-3. Snowpack Meltwater Production and Movement

As was pointed out earlier (see Figure 2-1), before snowmelt becomes runoff from a watershed, a number of processes occur. These processes involve a change in character of the snow crystals, changes in snowpack temperature and density, and the movement of meltwater through the pack. The changes in the internal energy of the snowpack are relatively small and are usually neglected in deep packs, where other energy components dominate. For shallow snowcovers, however, these phenomena become more important.

a. Character of the snowpack. The formation of the snowpack begins with the deposit of new-fallen snow of relatively low density (i.e., specific gravity). With time, the snowpack changes; the delicate crystals of snow become coarse grains, and the density of the pack increases. The metamorphosis from a loose, dry, and subfreezing snowpack of low density to a coarse, granular, and moist snowpack of high density is sometimes spoken of as "ripening" of the snowpack. A ripe snowpack is said to be "primed" to produce runoff when it becomes isothermal at 0 °C and its liquid-water-holding capacity has been reached. At this point, the only storage effect of the snowpack is that of "transitory" storage, resulting in a temporary delay of liquid water in transit through the pack. Although ripe snow is usually the relatively dense, coarse-grained snowpack characteristic of the spring, there is no restriction on the time of year that the snowpack may yield liquid water to the underlying ground surface. Midwinter rainfall or snowmelt may satisfy the "cold content" and liquid-water-holding capacity of the snowpack. After those deficiencies have been met, any further input of liquid water at that time will pass through the snowpack as drainage by gravitational

force. Figure 2-2 shows the features of a deep snowpack during a winter-spring season. Changes in depth, density, and snow temperature can be seen as the season progresses. Note the midwinter rainstorm in December where the snowpack became isothermal in only a few hours.

(1) Changes that take place within the snowpack are caused by several physical processes:

(a) Heat exchange at the snow surface.

(b) Percolation of meltwater or rain through the snowpack.

(c) Internal pressure attributable to the weight of the snow.

(d) Wind.

(e) Temperature and vapor pressure variation within the snowpack.

(f) Heat exchange at the ground surface.

As each new layer of snow is deposited, its upper surface is weathered by radiation, rain, and wind. The undersurface of the new layer may also be affected by ground heat. As a result, the snowpack is stratified, showing distinct layers and ice planes or lenses that separate individual snowstorm deposits. The interior of the pack is subjected to the action of percolating water and diffusing water vapor.

(2) During the melt season, on clear nights, a relatively shallow surface layer of the snowpack generally cools considerably below 0 °C, owing to the loss of heat to the sky by long-wave radiation; the liquid water may freeze in this layer to as much as 25.4 cm (10 in.) deep, but below this surface layer the liquid water remains unfrozen.

b. Drainage of snowmelt through the snowpack. Snowmelt moves through the snowpack vertically and horizontally. However, after the liquid-water conditioning of the snowpack has taken place, the movement of water through the pack is mostly straight downward to the ground/snow interface. Ice layers within the snowpack, however, tend to deflect the path intermittently, thereby resulting in an irregularly stepped pattern (see Figure 2-2). Analysis of meltwater movement through snow is more complicated than infiltration in a more static medium such as soil. The snowpack medium changes continuously as snow grains change in shape and size. In addition, as the snow melts and refreezes, impermeable ice layers form. Colbeck (1978) and Yosida (1973) have shown that as meltwater drains through the snowpack, there exists a wetting front that is isothermal at 0 °C and a lower layer in the snowpack below 0 °C. These wetting fronts may not be a uniform wave. Vertical flow fingers form around inhomogeneities in the snowpack (Marsh and Woo 1984). Because of these inhomogeneities in the snowpack, it is typically beyond the focus of operational snowmelt models to determine representation values of permeability and the effective porosity of the snow. The net storage effect on water draining through the snowpack is a time delay to runoff, on the order of 3 to 4 hr of storage time for moderately deep packs. In general, the time delay caused by transitory storage in the snowpack may be ignored when considering snowmelt or rainfall runoff from project basins whose areas exceed 518 or 777 sq km (200 or 300 square miles). This applies only to mountainous regions where slopes are adequate to ensure free horizontal drainage. Where horizontal drainage is inadequate (as in the Great Plains, in contrast to the mountainous region of the western United States), the delay to runoff caused by the snowpack may be much larger than for the vertical transit of water through the pack alone. Anderson (1973) has developed empirical relationships that represent drainage of snowmelt in the snowpack by using a time lag and attenuation.

2-4. Meltwater Infiltration

The ground conditions (both the soil mantle and underlying groundwater aquifers) are important in evaluating snowmelt runoff.

a. Unsaturated zone.

(1) The soil mantle functions as a reservoir, storing water, when available, to be used during periods when potential evapotranspiration exceeds current supply. In snow hydrology, there will be essentially no direct runoff until the soil storage is filled to its field capacity,

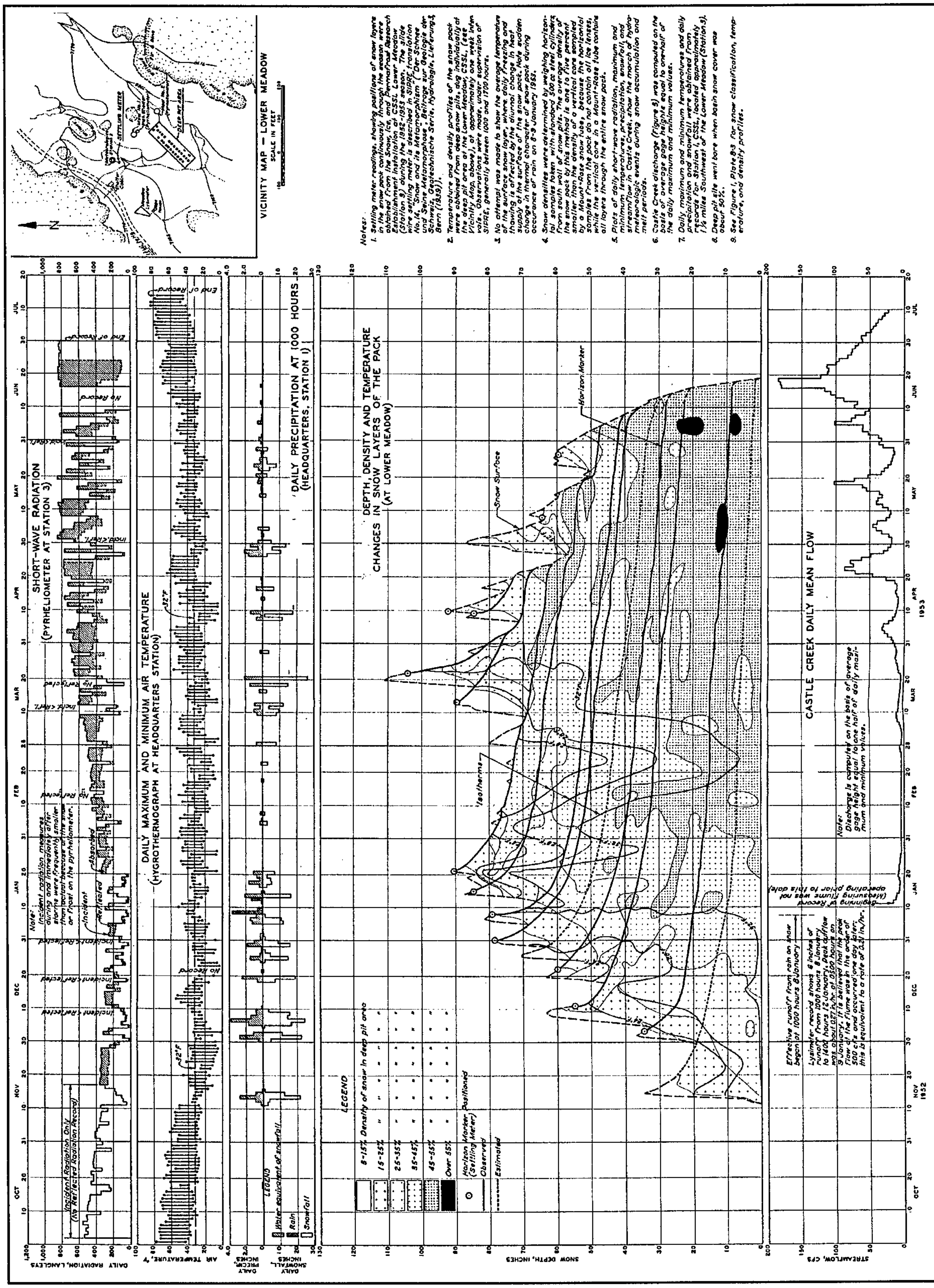

Figure 2-2. Snowpack characteristics (Plate 8-2, *Snow Hydrology*)

which is the total amount of water that can be held against gravity. The theoretical maximum capacity of the soil to withhold water permanently is determined by the difference between the "permanent wilting point" and the "field capacity" of the soil. For typical mountain soils, the maximum soil storage capacity ranges from approximately 10.2 to 20.3 cm (4 to 8 in.) of water in the zone from which stored water may be removed by transpiration or evaporation. After the field capacity of the soil has been reached, excess water may pass through the soil under gravitational force and appear later as subsurface or base flow components of streamflow. The time delay of transitory storage in the soil is integrated in the total basin storage effect.

(2) Direct measurements of soil moisture in project basins are generally lacking. While attempts are being made in some areas to obtain electrical resistance-type measurements of soil moisture beneath the snowpack, certain limitations currently restrict their use to qualitative interpretation. The principal difficulties are problems with calibration, lack of "buffer effect," inconsistency of results, disintegration of the sensing unit, and unrepresentativeness of individual samples. Accordingly, basin soil moisture conditions are generally estimated from indirect relationships involving earlier precipitation, duration of rainless days, groundwater levels, stream discharges, time of year, or other factors associated with soil moisture variation. For areas of deep snow accumulation, as in the mountains of the western United States, the soil moisture deficit is satisfied early in the snowmelt period, and in many areas it may often be satisfied in the fall from rainfall or snowmelt. In the latter case, the soil beneath the snowpack remains at or above the field capacity throughout the winter, and any loss by evapotranspiration will usually be supplied by winter snowmelt or rainfall. For years in which the soil moisture capacity is not filled by fall or winter rainfall or snowmelt, it is necessary to estimate the condition of the soil from preceding hydrometeorological events.

b. Frozen soil. Cases are known where losses are reduced significantly because of frozen ground, thus increasing runoff. The ground is generally unfrozen beneath deep mountain snowpacks because of the flow of heat from the ground, together with the insulating effect of the snowpack. Frozen ground will occur during winter or early spring, in areas where snowpacks are shallow, and where prolonged periods of subfreezing air temperatures prevail. Such conditions are characteristic of the northern Great Plains regions of the United States. Gray and Prowse (1992) state that the infiltrability of frozen ground is the single most important factor affecting the apportioning of snow water between direct runoff and soil waters in most northern regions.

(1) In general, the effect of frozen ground is to inhibit infiltration. In cases where the soil pores are small, liquid water entering the ground will refreeze within the surface layer and will retard further infiltration. Accordingly, the concept of satisfying soil moisture deficits for unfrozen soil would not apply and, in addition, the basin time delay for water in transit would be considerably reduced.

(2) The factors that affect the role of frozen ground in snow hydrology include frost types and hydraulic properties, changes in the routing of water through a watershed because of frozen ground, and the features of a streamflow hydrograph during a storm on frozen ground. Several structurally and hydrologically different types of frost may form when the soil freezes. The type of soil frost primarily depends on the moisture content at the onset of freezing and, to a lesser extent, the type of soil that freezes. Dingman (1975) found four types of frost commonly mentioned in the literature, those being concrete, granular, honeycomb, and stalactite. Only concrete and granular frosts occur in sufficient quantity or remain long enough to be of any hydrological significance. Concrete frost is most common in bare or sparsely covered soil and is the predominant frost type in soils frozen deeper than a few inches. Granular frost is most common in soils with higher organic material content and shallow freezing.

(3) The most important hydrological features of frozen ground are the change in the soil's permeability and, to a lesser extent, the volume of water bound in the frozen soil. Concrete frost is generally impermeable, although it may be interrupted by discontinuities in the soil surface. Frozen soil can also retain significant amounts of water in the soil column, particularly during the spring thaw when the subsurface frost layer holds meltwater above it (Alexeev et al. 1972).

(4) Frozen ground interferes with the normal path and time of travel of water through the watershed to the stream channel. The rate at which water infiltrates into the soil depends greatly on the conditions at the surface, as infiltration capacity is determined by soil type, soil moisture condition, and soil frost conditions. When the rate of water delivered to the soil exceeds its infiltration capacity, the excess water at the surface becomes overland flow. This surface water may also be stored in surface depressions until it can infiltrate, flow overland, or evaporate.

(5) The extent to which frost interrupts the normal routing processes of a watershed depends upon the extent of frozen ground. Dunne and Black (1971) found the areal extent of concrete frost in pastureland to be important to the timing of runoff. Discontinuities in concrete frost are common in forested areas, allowing more infiltration during frost conditions (Trimble, Sartz, and Pierce 1958). A general progression of frost occurrence by land-use type has been identified by several investigators (Storey 1955, Pierce 1956, Dingman 1975). The susceptibility of a land area to freezing is inversely related to the amount of ground cover and proportional to the degree of compaction of the soil. The general sequence of land-use types in accordance with their degree of frost susceptibility is bare cultivated ground, grassland, pasture, softwood stands, and hardwood stands. The proportions of these land-use types in a watershed influence the extent of frost, and, thus, the change in how the watershed responds.

(6) Water can run off frozen ground during rain on bare ground, rain on snow, and during snowmelts. In each of these events, the frozen ground effect depends on its extent at the event's beginning and the persistence of the frost throughout.

(7) The principal effects of frozen ground on the outflow hydrograph of the watershed are faster response with higher peak flow and greater volume in the total hydrograph. Simulations compared with hydrographs of actual storms over frozen ground show a distinct quickening of response and an increase in the peak outflow during the storm (Anderson 1978, Stokely 1980, Peaco 1981). The water's inability to enter into the soil reduces the amount of groundwater storage and increases the total volume of the hydrograph (Haupt 1967). Dingman (1975) suggests that a bimodal hydrograph could result if there is substantial frost melting during the storm. Upon the melting of the frost, overland flow could be reduced and infiltration increased, causing a dip in the hydrograph until the interflow and baseflow response appeared. In operational snowmelt runoff models, the effect of soil frost is accounted for in a soil moisture routine by controlling soil parameters and transfer coefficients. The state factor for frozen ground is usually a frost index. Examples of the use of a frost index are given by Anderson and Neuman (1984) and Molnau and Bissell (1983).

c. Saturated zone. Delay of runoff by ground and channel storage is a basic hydrological phenomenon. Direct evaluation of groundwater storage through the use of well records is impractical in mountainous areas because of the wide variability of conditions in a drainage basin. Streamflow-recession analysis is a way to indirectly evaluate basin storage. Volumes of water "generated" in a given period can be determined by use of standard recession analysis techniques.

2-5. Glacier Effects on Runoff

The presence of glaciers in a watershed or larger basin significantly affects runoff volume, frequency, and variability (Lawson 1993). Partial glacierization of a basin by as little as a few percent of cover can cause moderate to extreme variations in peak runoff magnitude and frequency over days, years, and decades. Runoff is not directly related to precipitation within a glacierized basin, so it is, at present, difficult to predict because of a lack of glaciohydrologic data and a limited, rather rudimentary knowledge of the glaciohydrologic processes controlling runoff.

a. Runoff from glacierized areas of a basin is generally greater than that from nonglacierized areas with similar precipitation, often by 3 to 10 times. The majority of runoff from ice-covered areas comes during the melt season, generally from mid-May to mid-September. At progressively higher latitudes or in higher elevation basins, the time of flow is reduced. Because glaciers act as natural storage reservoirs that retain a large proportion of the winter precipitation in their accumulation areas, they generally moderate annual streamflow. During warm, dry, and sunny

summers, water released from storage by melting of ice compensates for reduced runoff from precipitation. During cooler and wetter periods, the proportion of runoff from precipitation increases and supplements reduced meltwater runoff.

(1) Year-to-year variations in runoff also vary with percentage of glacier cover within the basin. Calculations of the coefficient of variation (CV) for annual runoff from partly glacierized basins suggest that there is minimum variability when ice covers between 35 and 45 percent of the basin area. The CV then is less than those of nearby nonglacierized basins, which tend to vary (and have a similar CV) as the precipitation totals vary. In addition, the CVs for monthly variations in runoff are lowest at the height of the melt season, but highest early in the season as the glacier's drainage system develops. In contrast, diurnal fluctuations in glacially fed rivers are greater than those in nonglacial rivers. They reflect primarily total energy input, which determines melt rates of the snowcover and glacier ice and, secondarily, the nature of the drainage system within the snowpack and glacier as it develops through the melt season.

(2) In addition, the timing of the peak diurnal discharge, as well as its magnitude, varies with the time of the melt season, occurring progressively earlier in the day with a larger magnitude and daily range later in the melt season. Peak seasonal flows are typically delayed compared with adjacent nonglacierized basins. For example, in the northwestern United States, discharge peaks in July or August, whereas it peaks in May in nonglacierized basins. This response reflects reduced albedos as the snow cover melts and maximum melt rates later in the year reflecting the minimal cloud cover and low precipitation of the region.

b. The effect of rainfall on runoff may differ from nonglacierized basins as well, reflecting the state of the drainage system within the snowpack and glacier. Early in the melt season, when drainage is incompletely developed, peak runoff from rainfall may lag significantly, whereas late in the season, when it is well developed, water movement through the glacier-covered area may be rapid and the response in runoff almost immediate. Flooding commonly follows heavy rainfalls after extended periods of high ice-melt rates, particularly when the drainage system is well-developed. Sudden, sometimes catastrophic, flooding also results from the unexpected release of water stored within or under the glacier or from drainage of ice-dammed lakes. Finally, snowfalls (rather than rain) at any time in the ablation season interrupt ice melt because of the reduced surface albedo, causing a decrease in runoff over several days or more.

c. The processes of snow metamorphosis and snowmelt upon the glacier follow those described elsewhere in this manual. A significant difference in defining snowmelt runoff, however, exists because the snowpack lies on glacier ice, which may be impermeable owing to seasonal freezing, and will therefore require warming or melting of internal passageways before runoff can actually take place. This process is not well-documented and its nature, the factors controlling it, and its rate cannot currently be defined. It is clear that runoff is significantly delayed because meltwater is stored within the snowpack above the ice. Only by surface drainage does the snowmelt slowly reach streams draining the ice-covered areas. Meltwater produced by ablation of glacier ice similarly is delayed from reaching basin streams by the early season absence of a well-connected drainage system inside and below the glacier. Therefore, while hydrometeorological analyses can be used to predict melting rates for the ice surface as it is gradually exposed by snowmelt, accurately predicting daily, monthly, or seasonal discharges remains elusive.

d. Drainage within (englacially) and below (subglacially) the glacier is inherently complex. In a general sense, water flows englacially either within the ice grains, eventually forming capillaries or small tubes that intersect larger ice-walled conduits within the glacier, or through larger drainage features, including crevasses, fractures, and moulins, that intersect or feed conduits at depth within the ice. The ice-walled conduits progressively join larger conduits at depth, forming an upward branching network. Once at the bed, water moves in conduits and cavities incised into the bed or into the overlying ice. These conduits are tributaries to larger tunnels discharging at the ice margin. Some water also flows as a thin film at the ice/bed interface or as groundwater in subglacial sediments. However, the configuration, distribution,

and variability of this drainage system are incompletely described and are generally speculative.

e. Conceptual models specific to glacierized basins attempt to predict runoff by separating the procedure into two steps: one to calculate meltwater production and the other to calculate drainage, both from the glacierized and nonglacierized portions of the basin. Meltwater production is simulated by considering the physical processes and their effect on melt rate. Drainage from the glacierized part of the basin, however, is poorly simulated by existing models, mainly because of the lack of empirical data and lack of a sufficient understanding of the processes controlling flow rates and volume, and water storage. In some models, a linear reservoir with a retention time built in is used to simulate glacier drainage. In others, the glacier is considered an extremely thick snowpack and treated as such. Operational physical models that may or may not treat glacier drainage and storage separately include those of Anderson (1973) as modified by Nibler (cited in Fountain and Tangborn 1985), Baker et al. (1982), Lang (1980), Lang and Dayer (1985), Tangborn (1984, 1986), and those in Power (1985). In addition, none of the models is strictly physically based, but incorporate statistical treatments where process relationships are unknown. These conceptual models illustrate the present approaches to glacierized basin runoff predictions. In general, none of the operational models accurately predict the frequency and magnitude of peak, seasonal, and annual flows, and each of the models is basin-specific. Only Lang and Dayer (1985) apply their model to hourly and daily predictions of runoff; overall, their model best predicts seasonal runoff for an operational scheme.

Chapter 3
Collection and Analysis of Basic Data

3-1. General Introduction

Knowledge of snowfall amounts, the amount of snow accumulation on the ground (snowcover), and their spatial distribution throughout the watershed or basin area of interest is essential for the effective use of snowmelt runoff models. Thus, operational snowmelt forecasting programs must include activities to measure or acquire accurate snowfall and snowcover data.

a. Goodison, Ferguson, and McKay (1981) define snowfall as "the depth of fresh snow which falls during a given 'recent' period.....a single storm, a day, a month or a year." They also define snowcover as "the amount of snow on the ground at the time of an observation." They note that, "The ground may be either completely or partly covered." The amount of snowfall and snowcover is influenced by many variables and, thus, typically can vary substantially over even relatively small areas. Variation over regions can be great.

b. The accurate measurement of snowfall and snowcover at a given point and of the spatial distribution of snowfall and snowcover over the basin is difficult and can consume the resources that are available to operational snowmelt forecasting programs. The parameters that are measured to define snowfall and snowcover are snow depth, snow water equivalent (water content), snow density and location, and extent of the snowcover. Table 3-1 summarizes the techniques that are available to measure snowfall and snowcover. Some methods allow for the measurement of snowfall and snowcover at a point, while others are adapted to the measurement of the areal extent of the snowcover.

3-2. Summary of Snow and Snowcover Parameters

a. Snow depth. Snow depth is routinely measured using graduated snow rulers that are installed to the ground surface. In recent years, snow depths have been successfully measured with acoustic snow depth sensors, which can be interfaced to remote data-collection systems. These sensors employ ultrasound range finders that measure the distance from a fixed elevation above the snowpack to the snowpack surface. The sensor is installed at an elevation greater than the highest expected snow depth before snowfall and the baseline is electrically set in the transducer or data-collection system. The acoustic snow depth system can operate with a ±1-cm accuracy (Metcalfe, Wilson, and Goodison 1987). The accuracy of snowfall measurements with snow rulers, snow boards, and snow gauges are affected by siting conditions and observer bias. Thus, at each observation station, multiple measurements are usually made to acquire a representative depth measurement. The literature documents the effects of siting or exposure on the accuracy of snowfall measurements. Peck (1972), Goodison (1978a), and Larson and Peck (1974) discuss the proper siting of measurements that minimize the local effects of drifting. Snow depth is usually expressed in inches or centimeters.

b. Snow water equivalent (water content). Snow water equivalent (SWE) is defined as the equivalent depth of water in the snow that is sampled and is normally expressed in inches or centimeters of water. The water content of either newly fallen snow or of the accumulated snowpack has been traditionally measured by weighing a vertical core taken through the snowpack. This measure is the basis of snow surveys, which are conducted throughout the United States to obtain the spatial distribution of SWE in a watershed or region. Measurement of SWE is subject to a variety of errors (Work et al. 1965, Goodison 1978b). Probably the most common error results from the field acquisition of an incomplete core of snow in the sampler tube. This may be caused by clogging of the cutter by corky snow, obstructions such as stones or sticks, or sticking of the snow to the tube. Such things can generally be detected by comparing the length of core with depth of the snow at the time the core is taken. Another source of error is the sampling of ponded water in the lower portion of the core, resulting from poor snowpack drainage. In such cases, the water equivalent may be computed by multiplying the depth of snow by densities obtained at nearby sample points. Any dirt or other foreign matter must be removed from the cutter end of the sample before the core is weighed. At sites where frequent observations are made, care must be exercised to avoid holes

Table 3-1
Snowfall and Snowcover Measurement Techniques

Measurement Class	Method Name	Application	Parameter Measured	Characteristics of Method
Simple linear measurement	Graduated snow ruler	Fresh snowfall	Depth	a. Point measurement; representativeness of measurement location of concern. b. Preparation of measurement site needed for each new snow event.
		Accumulated snowpack	Depth	a. Point measurement; representativeness of measurement location of concern. b. Measurement frequency a function of personnel availability.
	Snow board	Fresh snowfall	Depth	a. Point measurement; representativeness of measurement location of concern. b. Preparation of measurement site needed for each snow event.
Gravimetric	Precipitation gauges a. Nonrecording bucket gauge	Fresh snowfall Accumulated snowfall	Water equivalent, in. Water equivalent, in.	a. Point measurement; representativeness of measurement location a concern. b. Capture efficiency a function of gauge baffling and local wind regimes. c. Preparation of gauge needed for each new event.
	b. Recording weighing/tipping bucket gauges	Accumulated snowfall	Water equivalent, in. Snowfall rate, in./hr	a. Point measurement; representativeness of measurement location a concern. b. Capture efficiency a function of gauge baffling and local wind regimes. c. Can provide a continuous record. d. Gauge maintenance relatively infrequent depending on chart life and bucket capacity.
	c. Electronic balance	Accumulated snowfall Snowfall rate	Water equivalent, in. Snowfall rate, in./hr	a. As per recording precipitation gauges. b. Can provide rapid response times.
	Snow samplers (snow tubes)	Accumulated snowfall (snowpack)	Depth Water equivalent, in.	a. Point measurement; representativeness of measurement location a concern. b. Measurement frequency a function of personnel availability.
	Snow pillows and snow triangles	Accumulated snowfall (snowpack)	Water equivalent, in.	a. Point measurement; representativeness of measurement location a concern. b. Can be adversely affected by 'bridging' caused by ice lenses. c. Large and bulky; installation difficult. d. Can provide a continuous record.
Calorimetric	Freezing, alcohol solution or dilution calorimetric methods	Any snow sample	Liquid water content of snow sample (weight basis)	a. Point measurement; representativeness of measurement location a concern. b. Requires careful sample management prior to analysis c. Analysis relatively complex. d. Frequency of analysis determined by personnel availability.

Table 3-1 (continued)

Measurement Class	Method Name	Application	Parameter Measured	Characteristics of Method
Electromagnetic A. In situ sensors	a. Gamma radiometers	Accumulated snowpack	Water equivalent, in.	a. Point measurement; representativeness of measurement location of concern. b. Gamma radiation can be harmful to health. Cannot be left unattended in field. c. Can provide density profiles of snowpack.
	b. Acoustic sensors	Accumulated snowpack	Depth	a. Point measurement; representativeness of measurement location of concern. b. Setup, installation needs calibration. c. Can provide a continuous record with frequent readings. d. Adaptable to automatic data capture.
	c. Optical snow gauge (transmissiometer)	Snowfall	Snowfall rate, in. Snowfall mass conc., g/cc	a. Point measurement; representativeness of measurement location of concern. b. Output a function of snow particle size and crystal type and fall velocity. c. Can provide a continuous record with instantaneous readings. d. Adaptable to automatic data capture.
B. Remote sensors (satellite or airborne mounted)				GENERAL: Remote sensor data generally do not represent point measurement but rather are applicable to wide-area surveys. Depending on data use resolution of the sensor data may be of concern. Except for visible photos, data are acquired in digital formats, and thus efficient input to automated data systems is possible.
	a. Natural terrestrial gamma radiation	Accumulated snowpack	Snowpack extent Water equivalent, in.	a. Background gamma radiation survey must be winter flight survey. b. Data are amenable to automated data analysis systems. c. Groundtruthing survey needed.
	b. Visible photography	Accumulated snowpack	Snowpack extent	a. Weather and clouds interfere with data acquisition. b. Computerized data systems require analog data to be digitized before use.
	c. Microwave	Accumulated snowpack	Snowpack extent Water equivalent, in.	a. Weather and clouds can interfere with data acquisition. b. Groundtruth information desirable.
	d. Radar-accumulated snowpack	Snowpack extent	Depth	a. Data acquisition possible in presence of clouds and certain weather conditions.
	e. Multispectral images	Accumulated snowpack	Snowpack extent	a. Weather and clouds can interfere with data.

left by prior sampling. Observer blunders such as misreading the snow depth or sampler weighing scales also happen. A comprehensive discussion of the methods used to take snow core measurements is available.

c. Snow density. Snow density is defined as the weight of snow per unit volume of snow and has the units of pounds/cubic foot or grams/cubic meter. Snow density is obtained by dividing the SWE by the depth of snow as would be measured when taking snow core readings or by simply weighing a known volume of snow.

d. Areal extent of snowcover. The location and extent of snowcovers are usually estimated using remotely sensed data that can discriminate between snowcover and no snowcover. Snowcover extent is often expressed as a percentage or fraction of the total drainage area of interest that is covered by accumulated snow. In a number of snowmelt runoff models, it is desirable to know snowcover extent within a number of defined elevational zones in the watershed area.

3-3. Measurement of Snowfall and Precipitation

a. Snowfall depths. Snowfall is measured at a point using a snow ruler or snow board, limited-capacity nonrecording snow gauges, recording-weighing-type precipitation gauges, or high-capacity precipitation-storage gauges. The depth of snow that has fallen during some recent period can be measured with a graduated rule (snow ruler). Snowfall depths are sometimes measured on a snow board whose surface has been kept free of snow before the snowfall. The water equivalent of the newly fallen snow can be estimated knowing snow density, or the snow can be melted in samples taken from on top of the snow board.

b. Nonrecording precipitation gauges. Nonrecording snow gauges have been used extensively to measure snowfall water equivalent. In Canada and the United States, this type of gauge has been designated as the official instrument for measuring snowfall water equivalent. The MSC Nipher shielded snow gauge (Goodison 1978a) is the Canadian standard, whereas the NWS alter-shielded 20.3 cm (8-in.) standard gauge (Larson and Peck 1974) is the United States standard. Nonrecording snow gauges need to be emptied frequently, usually once a day. The accumulated snow is melted and either weighed or measured in a glass that is graduated to obtain the water equivalent.

c. Recording precipitation gauges. Recording-weighing type of precipitation gauges measure both solid and liquid precipitation. In these gauges is a simple spring balance, whose mechanical displacement is recorded on a chart or converted to an analog electrical output. These outputs can be recorded onsite or telemetered by telephone, radio, or satellite (Metcalfe, Wilson, and Goodison 1987). The Universal and Fisher Porter are examples of the recording-weighing-type of precipitation gauges. The capacities of these gauges are 300- to 600-mm water equivalent, and the collection orifice is 20.3 cm (8 in.) in diameter. In cold climates, these gauges require an antifreeze charge, typically ethylene glycol, to prevent freezing in the collector. In addition, a layer of light oil is added to prevent evaporation. Alter shields are suggested for these gauges to reduce wind effects on collection efficiency. Recording-weighing type of precipitation gauges need to be visited periodically to check calibration, to empty the storage devices when capacity is reached, to change charts, and to replenish and mix antifreeze and oil. Large-capacity storage gauges, up to 2540 mm of water equivalent, are used at remote or unattended sites, such as mountainous regions characterized by high precipitation. Antifreeze, oil, and an alter shield are suggested with these. They can be automated for telemetry by connecting the storage gauge to a float and stilling well in an adjacent shelter.

d. Measurement errors. The effects of wind on gauge catch have been reported by Peck (1972), Goodison (1978a), and Larson and Peck (1974). Methodologies have been developed for adjusting the measured gauge catch to account for the effects of different meteorological variables (Hamon 1973, Rawls et al. 1975).

3-4. On-ground Measurement of Snowcover

a. Snow pillow. The snow pillow is a nondestructive technique for measuring the SWE of the snowcover. The snow pillow has been used extensively in the western United States, most notably by the U.S. Natural Resources Conservation Service in their SNOTEL network (Crook 1986). Snow pillows are constructed with various shapes, sizes, and materials; they are fluid-filled pillows in which fluid pressure responds to the weight of snow that is lying on them. The pressure of the fluid in the pillow is measured with a manometer or pressure transducer, which may be interfaced to a digital data-collection and transmission system. The pillows are made from butyl rubber, neoprene rubber, sheet metal, or stainless steel. Discussions of differing types of pillows and the specifics of design and operation are presented by Davis (1973) and Cox et al. (1978).

(1) Pangburn and McKim (1984) discussed a potential snow triangle to avoid the hydraulic problems associated with fluid-filled pillows. The snow triangle replaces the fluid-filled pillow with a plywood triangle having an area of 1.5 m^2. The plywood triangle is placed on three load cells that provide an electrical output proportional to the weight of snow on the triangle.

(2) Both snow pillows and snow triangles are affected by bridging caused by ice lenses forming in the snow pack. This bridging stops the pillow or plywood triangle from sensing the full weight of the overlying snow so that there is decreased or lagged detection of SWE. Snow pillows can be an effective instrument for monitoring SWE where formation of ice lenses is not prevalent, such as in shallow snow packs or in deep mountainous packs in the western United States.

b. Radioisotopic gauges. Radioisotopic gauges have been used to make measurements of snowpack water equivalent at remote, unattended sites and to transmit these data to a central receiving station. These gauges depend on the fact that the water in the snowpack attenuates any gamma radiation that is emitted by any source under the snowpack. The intensity of the radiation received by a detector above the snow surface is related to the snowpack's total water equivalent, provided background radiation levels are known. One of the first radioisotopic gauges was developed by USACE in 1955. The USACE gauge consisted of a cobalt 60 gamma ray source placed at the ground surface and a Geiger-Muller radiation-detector type (G-M tube) suspended 4.6 m (15 ft) overhead. Since this first development, many systems have been tested. More recently, radioisotope gauges have become portable (Young 1976) and use naturally occurring uranium as a source (Morrison 1976). They have been used to profile SWE and density (Smith, Halverson, and Jones 1972). Care must be exercised when making these measurements to avoid inappropriate radiation exposure to operating personnel. In addition to artificial radiation sources, natural radiative emissions from elements in the soil can be used to measure SWE at a point.

c. Snow surveys. The common practice for making snow surveys is to sample and measure the snow water equivalent at a number of points along an established line called a snow course. Snow courses are located with the objective of obtaining data representative of a given area, the number of samples depending largely upon the terrain and meteorological characteristics of the area. Other factors such as accessibility, availability of funds, and purpose for which the data are to be used, must, of course, be considered in the establishment of the network of sampling stations.

(1) Selection for a snow course site should be based on the same general requirements as for precipitation gauges, with the following being considered:

(a) Meteorological conditions with respect to storm experience.

(b) Position with respect to large-scale topographic features.

(c) Position with regard to local environmental features, such as exposure, aspect, orientation, and ground slope.

(d) Conditions on the site itself, such as local drainage and the occurrence of brush and rocks.

In addition, snow courses should be located to adequately sample ranges in elevation, and they also should be so located that they are representative of average basin melt conditions, as well as basin snow accumulation. As is the case for precipitation gauges, snow courses should be located in areas well protected from wind, since wind erosion and drifting snow cause unrepresentative snow accumulations. An ideal location would be an opening in the forest surrounded by hills for protection from high winds and sloped sufficiently to permit runoff of water beneath the snowpack. The number of sample points is variable, depending largely upon the consistency of the distribution of snow. Sample points are located with the objective of avoiding variations in snow depth from causes such as drifting, interception by trees, and the presence of boulders or other obstructions. If protection from wind is altogether lacking, the sampling points must be spread over a wide area to average out variations caused by drifting.

(2) In general, five sample points are probably adequate for well-located snow courses upon which there is a minimum of irregularities caused by drifting or wind erosion, if the ground surface is smooth and clear of all obstructions, and if the snow course is not too close to the forest or other local obstructions to be influenced by local irregularities in deposition. When conditions are less than ideal, however, additional snow course points are required to adequately sample the water equivalent.

(3) Although care is exercised in selecting locations having stable physical features, there may be changes affecting the deposition of snow at sampling points. A common change in physical features is the removal of all or a portion of the surrounding timber by fire, cutting, bug infestation, or severe wind storms. On the other hand, an opposite effect can be produced by the growth of brush or timber in the vicinity of the sampling points. In the latter case, annual changes may not be detectable; nevertheless, the change over a period of years may be significant. Another important effect of physical changes is improper drainage of free water as a result of obstructions such as beaver dams or accumulation of debris in drainage channels in the snow-course area. Occasionally, physical features may change sufficiently to necessitate abandonment of the snow course. Often, however, the location is acceptable despite some changes in physical features. In such cases records must be adjusted.

(4) Basic data from snow courses are obtained under cooperative arrangements among various Federal, State and private organizations. The many details pertaining to snow surveying for obtaining the water equivalent of the snowpack at a given point are beyond the scope of this manual; for details see the comprehensive reports on snow surveying by U.S. Soil Conservation Service (1972), Atmospheric Environment Service (1973a,b), and World Meteorological Organization (1974).

3-5. Remotely Sensed Measurement of Snowcover

a. Aircraft measurements. Aircraft measurements have been used historically to define the spatial distribution of snowpacks, especially in inaccessible, remote areas where point snow-course measurements could not be obtained. Smith, Cooper, and Chapman (1967) found that measuring the distribution of snow by aerial photography was a practical methodology for areas of complex relief, and that snow depth could be determined in such areas with high precision. Others have used aerial overflights for determining snowline elevation—for example, in the Columbia River Basin by both the USACE and British Columbia Hydro and Power Authority. Aircraft surveys can be an effective method of gathering data on snow depth and snowline; however, such surveys are limited to suitable flying conditions and can be relatively expensive and time-consuming (Rango 1977; Goodison, Ferguson, and McKay 1981).

b. Airborne gamma survey. As previously mentioned, the water contained in snowpack attenuates gamma radiation. Natural terrestrial gamma radiation is emitted from the potassium, uranium, and thorium radioisotopes in the upper 20 cm of soil. The levels of this natural terrestrial gamma radiation are monitored using sensors in a low-flying aircraft (150-m altitude). When adjusted for background, the intensity of the radiation can be can be related to SWE. Terrestrial gamma surveys are conducted before snowfall to obtain background readings (Bissell and Peck 1973, Loijens 1975).

(1) Airborne gamma survey technology was originally developed in the USSR (Russia) in the 1960s (Kogan et al. 1965) and has developed into a fully operational tool for the U.S. National Weather Service (NWS) (Carroll and Allen 1988). Mean areal SWE can be obtained with a root mean square error of less than 1.25 cm by calculating the difference between measurements made over bare ground and snow-covered ground. The accuracy of this method is affected by many things, including changing soil-moisture conditions and radon gas (Vadnais 1984; Carroll and Jones 1982, 1983). The great advantage of aerial gamma surveys is their large-area coverage, which minimizes the effect of high local variation.

(2) The NWS has developed a National program that includes two terrestrial gamma radiation systems on low-flying aircraft over a network of more than 1600 flight lines covering portions of 25 U.S. States and 7 Canadian Provinces. The limitations of airborne gamma surveys are their restrictions to relatively flat areas and the precise navigation needed to correlate to groundtruth data.

c. Satellite observations. Remote sensing of snowcover using satellites has been studied since the 1960s and used most successfully for delineating snow-covered areas. To date, however, there are no operational automatic snowcover mapping algorithms. Historically, the two principal satellite systems used for snowcover delineation in the United States have been the LANDSAT and Advanced Very High Resolution Radiometer (AVHRR) systems. Imagery from LANDSAT has a swath width of 185 km, a pixel size of 30 m, and a return interval of 16 days. Cloud cover in imagery may lengthen the period between usable images. Moreover, the tradeoff between large area coverage and high spatial resolution of the LANDSAT imagery is the large data volumes. A single LANDSAT scene contains over 200 megabytes of data, which makes analysis of large river basins, covered by several images, or time sequences of images, difficult without workstation-level or better computers. LANDSAT data have proven useful for the study of medium to small river basins.

(1) Snow mapping from satellites has developed mainly since the 1970s. Rango and Itten (1976) used both supervised and unsupervised computer classification techniques to map snow-covered area from LANDSAT MSS data. Snow in trees and melt-freeze snow were classified, but the criteria were not specified. Martinec and Rango (1981) used LANDSAT MSS data to estimate the distribution of SWE over an alpine basin. Dozier (1989) addressed the problem of calculating snow reflectance from LANDSAT Thematic Mapper data and used the difference in reflectance in Bands 2 and 5 to discriminate snow from clouds and bare ground. Crane and Anderson (1984) used Defense Meteorological Satellite Program (DMSP) data to discriminate clouds from snow and water clouds from ice clouds. Baumgartner, Seidel, and Martinec (1987) demonstrated the supplementation of LANDSAT data (high spatial resolution) with AVHRR data (high temporal resolution) for improved snowcover depletion estimates. An automatic mapping method was developed by Dozier and Marks (1987) for using the information in digital elevation models without requiring precise registration of the images to the models. This method required the use of an atmospheric transmission model and the knowledge of grain sizes and contaminants. Dozier (1989) demonstrated automatic snow mapping based on apparent planetary (spectral) reflectance. Thresholding and normalized difference ratios for Bands 1, 2, and 5 were used to identify snow in shadow and to discriminate sunlit rocks, soils, and clouds from sunlit snow. AVHRR data were used by Baglio and Holroyd (1989), registered to a digital elevation model, to test an interactive snow-mapping system.

(2) In an operational setting, image data from the AVHRR on the National Oceanic and Atmospheric Administration (NOAA) polar orbiting satellites and image data from the Geostationary Operational Environmental Satellite (GOES) are used for snowcover mapping. The resolution of AVHRR images is about 1 km, making the data usable for large river basins and regional coverage. These data are collected and disseminated by the NWS National Hydrologic Remote Sensing Center (NWS 1992) and provide daily maps of the percentage of snow cover in approximately five elevation bands for each of more than 500 major river basins in the western United States and Alaska. The NWS also provides complete coverage of the United States and Canada, with additional basin boundary sets to map snow cover for

the upper Midwest, the Great Lakes, New England, and Canada. These data are electronically accessible to end-users, in near real-time.

(3) Passive and active microwave sensors have been used for snowcover measurements and can operate in all weather conditions. Since 1978 the NIMBUS-7 satellite has provided data from the Scanning Multi-Channel Microwave Radiometer (SMMR) with a resolution of 30 km^2. SMMR is a five-frequency, dual polarized instrument that measures the upwelling microwave radiation from 6.6 to 37.0 GHz (Gloerson and Barath 1977). Goodison and Walker (1993) have found that SMMR data were sufficient to measure snow extent and SWE, when the snow was dry, in the Canadian prairie where ground measurement stations are sparsely located. Moreover, time-sequential data have shown the promise in the detection of wet snow. Others have shown the utility of passive microwave SMMR data to map snowcover properties over relatively flat homogeneous areas like the Canadian prairies (e.g., Chang, Foster, and Hall 1987; Rango, Chang, and Foster 1979), and at large scales, the maps compare well with the NWS product. In 1987 the DMSP launched another microwave radiometer, the Special Sensor Microwave Imager (SSM/I). The SSM/I is a four-frequency dual polarized radiometer that operates in the frequency range of 19.3 to 85.5 GHz. Measurements from this instrument have been shown to be useful for mapping snowcover extent, and algorithms to recover SWE are being developed. The goal for disseminating these data is near real-time for operational use. The main problems with passive microwave data are coarse resolution and lack of any general algorithm for estimating SWE that works in areas that are not large, flat, and homogeneous.

(4) The future looks promising for using remote-sensing inputs for operational snow hydrology models, as well as for research. Automatic classification algorithms for mapping snowcover extent and SWE appear to be forthcoming within the next decade. While no single sensor system or platform currently offers real-time frequent measurement of even snow-covered areas, the rapid evolution of remote-sensing and computing technologies, including geographic data systems, will allow the merging of sensor data sets (orbital, airborne, and ground based), which should improve operational forecasts substantially.

3-6. Snow Analysis

In addition to snowcover measurements, other hydrometeorological data are required for snowmelt simulation and hydrologic forecasting. These variables include air temperature and precipitation at a minimum; however, if energy budget methods are employed, such variables as wind speed, dew point, solar radiation, and others would need to be available. Table 3-2 summarizes the required data types, along with comments on their purposes and applications. Besides the data requirements described in Table 3-2 for snow analysis, physical data on standard streamflow measurements and watershed characteristics must be obtained. The application of a snowmelt simulation model typically takes the path of calibrating the transformation models in warm (nonsnowmelt) conditions and then calibrating the snowmelt routine to input the melt for the transformation model for the accumulation-ablation period. To do this the hydrologist requires long-term continuous discharge records (preferably greater than 10 years) for calibration and validation. The physical data, such as area-elevation data, type and density of land cover, slope and aspect of watershed elements, are of prime importance in mountainous areas. This is particularly important for distributed systems that compute snowmelt based on distinctly defined zones of elevation, subwatersheds, or hydrologic response units (HRUs).

Table 3-2
Data Requirements for Snow Analysis

Data Type	Physical Element or Purpose	Application
Streamflow (*Q*)	a. Continuous discharge b. Runoff volumes	a. Hydrograph analysis, model calibration b. Water supply analysis, forecasting
Precipitation (*P*)	a. Basin moisture input b. Estimate of SWE	a. Hydrograph analysis, model calibration b. Water supply forecasting
Air temperature T_a	a. Rain/freeze interface b. Index to all energy exchanges c. Factor in energy budget estimates	a. Modeling snow accumulation b. Modeling snowmelt (temp . index) c. Modeling snowmelt (energy budget)
Snow water equivalent (SWE)	a. Estimate of precipitation b. Index to basin water supply c. Snowpack quantity during ablation	a. Analysis, model calibration b. Water supply forecasting c. Modeling snowmelt
Areal snow cover	a. Extent of basin snow cover b. Snowline elevation	a. Model calibration b. Parameter in forecast models
Snowfall	a. Estimate of SWE, precipitation b. Accumulation of snow	a. SWE, precipitation applications b. Avalanche forecasting
Snow density	a. Estimate of SWE, precipitation b. Condition of snow	a. SWE applications b. Avalanche conditions, snow loads
Snow depth	a. Estimate of SWE, precipitation b. Estimate of weight	a. SWE, precipitation applications b. Snow load on structures
Snow albedo	Solar energy absorption	Modeling (energy budget), design floods
Solar radiation	Solar energy flux	Modeling (energy budget), design floods
Wind velocity (*v*)	Estimate of convection/condensation energy flux	Modeling (energy budget), design floods
Dewpoint temperature T_d	Factor in estimate of condensation energy flux	Modeling (energy budget), design floods

Chapter 4
Snow Accumulation and Distribution

4-1. General

A necessary ingredient in snow runoff analysis is determining the quantity and distribution of snow—more specifically the SWE—that exists in the basin prior to the onset of runoff. The SWE will be the primary determinant governing the magnitude of the snowmelt runoff volume; and the distribution of the snowpack in the basin (whether it be at low or high elevations) will be a factor in determining the rate of melt during the melt season. The SWE estimate must either directly or indirectly consider the process of snow accumulation and distribution, which involves a variety of meteorological and topographical interactions in the basin during the winter accumulation period. This process is much more complex than a rain-only situation, since temperature and elevation play such a prominent role in determining whether precipitation falls as rain or snow. The choice of methodology to determine snow accumulation depends upon data availability, the amount of effort to be expended, and the type of application involved. This chapter will describe alternative approaches for both analysis and forecasting, ranging from simple estimates of a single basinwide average to the detailed simulation of snow accumulation using a continuous model.

4-2. Precipitation, Snowfall, and Snow Accumulation

In the middle latitudes, precipitation usually falls as a result of the colloidal instability of a mixed water-ice cloud at temperatures below 0 °C (32 °F). Snow and rain forms in the atmosphere through a dynamic process. Winter precipitation begins as snow crystals in subfreezing portions of clouds. As the snowflakes fall through the atmosphere, they later melt into raindrops when they fall through warmer, above-freezing air at lower elevations. The melting level air temperature for snowflakes falling through the atmosphere varies from 0 to 4 °C (32 to 39 °F), but it is usually about 1-2 °C (34-35 °F). Accordingly, on the Earth's surface, snow falls at elevations higher than the melting level, while rain falls at elevations lower than the melting level. Figure 4-1 shows the frequency of observed forms of precipitation at

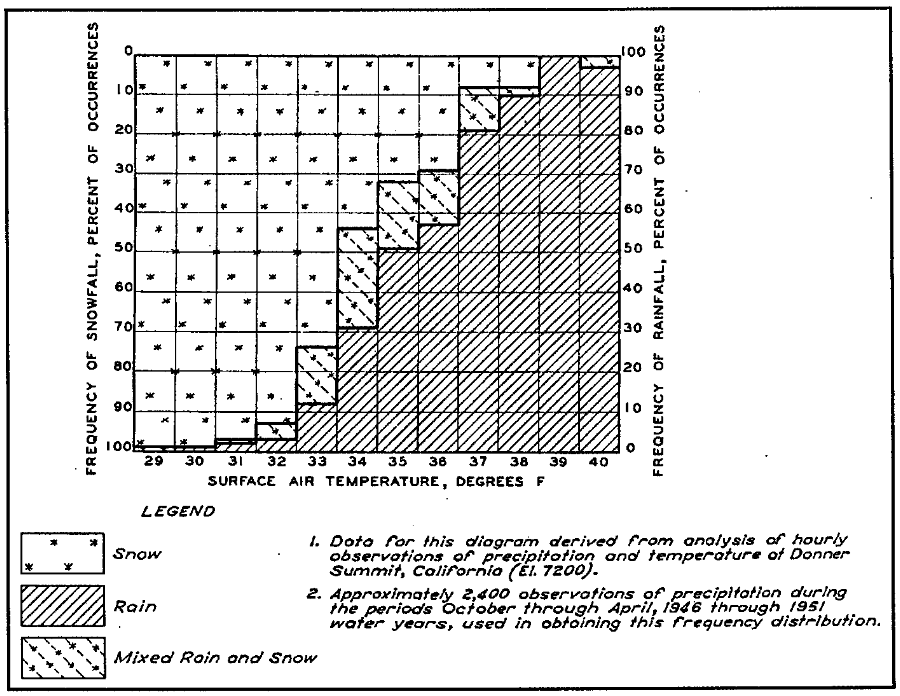

Figure 4-1. Forms of precipitation versus temperature (Figure 1, Plate 3-1, *Snow Hydrology*)

Donner Summit, California. The most significant thing that determines rain or snow is the elevation of the melting level. This is particularly important in mountainous regions. Factors influencing the amount and distribution of precipitation in the form of snow and the SWE may be classified as being meteorological and topographical. Meteorological factors include air temperature, wind, precipitable water, atmospheric circulation patterns, frontal activity, lapse rate, and stability of the air mass. Topographical factors include elevation, slope, aspect, exposure, and vegetation cover.

4-3. Watershed Definition

a. Overview. There are two basic approaches for defining a computer model of a watershed and, therefore, the distribution of snow in that model. A lumped model assumes that the progression of each variable through time (e.g., rain, snow, and soil moisture) can be reduced to a single computational algorithm that represents the entire basin. This is a considerably simplifying assumption in basins that have a wide variety of physical features, but such a model may produce satisfactory results for many applications. In a distributed model, the watershed is divided into subunits with variables being computed separately for each. The output from each subunit is combined to produce total basin output. Lumped models are generally limited to event-type modeling, where the model does not operate beyond a single runoff event. The distributed model formulation is required for continuous simulation, in which the model operates through low-flow periods by simulating the effects of evapotranspiration losses, groundwater, and other variables not normally of importance over short periods of flood runoff. Distributed, continuous simulation is being used more in recent years for both analysis and forecasting because of improved computer and data technology.

b. Lumped formulation. In this approach the basin's precipitation and snowmelt input is a single basin-mean quantity that is transformed to runoff by use of a unit hydrograph or similar methodology. Since this approach is normally limited to modeling runoff events only, the SWE prior to runoff must be determined indirectly and a single basin-average value provided as input to the transformation model. The SWE can be determined before the transformation model is executed, either with a separate computer program or perhaps by a manual estimate. Examples of using a lumped formulation in a snow environment might be as follows.

(1) A design flood derivation, in which the initial SWE is calculated in a relatively detailed but entirely independent analysis, using regression and frequency techniques. During melt, a single, basin-average value is acted upon by a depletion curve method discussed in Chapter 8.

(2) A rain-on-snow forecasting situation, in which rain dominates, but snowmelt can nevertheless add significantly to runoff. A single SWE value and snowline elevation is estimated by the forecaster, based upon a snow gauge located in the basin. With the rainstorm lasting only a few hours, the snowcover can be assumed constant during the melt computation.

c. Distributed formulation. For more detailed modeling of snow, a distributed definition of the basin is needed. This enables the snow accumulation process to be modeled directly, using continuous simulation, and it permits a more detailed accounting of snow during snowmelt. The oldest and currently most common approach in the distributed basin formulation is to subdivide the basin into zones or bands based upon elevation. (Technically, this type of formulation would still be lumped spatially.) On each elevation band, precipitation, snow, soil moisture, etc., are simulated independently; then moisture output from each band is totaled to obtain input into the runoff transformation routine. This method of subdividing the basin is a logical one, since in mountainous areas geographical, hydrological, and meteorological conditions are typically related to elevation. The snowband formulation is shown in Figure 4-2. The snowband method is available in several existing models. Setting up and configuring a basin model with these programs typically employs simplifying assumptions and generalized relationships, making the watershed definition a relatively easy process considering the amount of detail in the basic methodology. The snowband formulation is available in hydrologic models such as Hydrologic Engineering Center-1 (HEC-1) (USACE 1990) and Streamflow Simulation and Reservoir Regulation (SSARR) (USACE 1991).

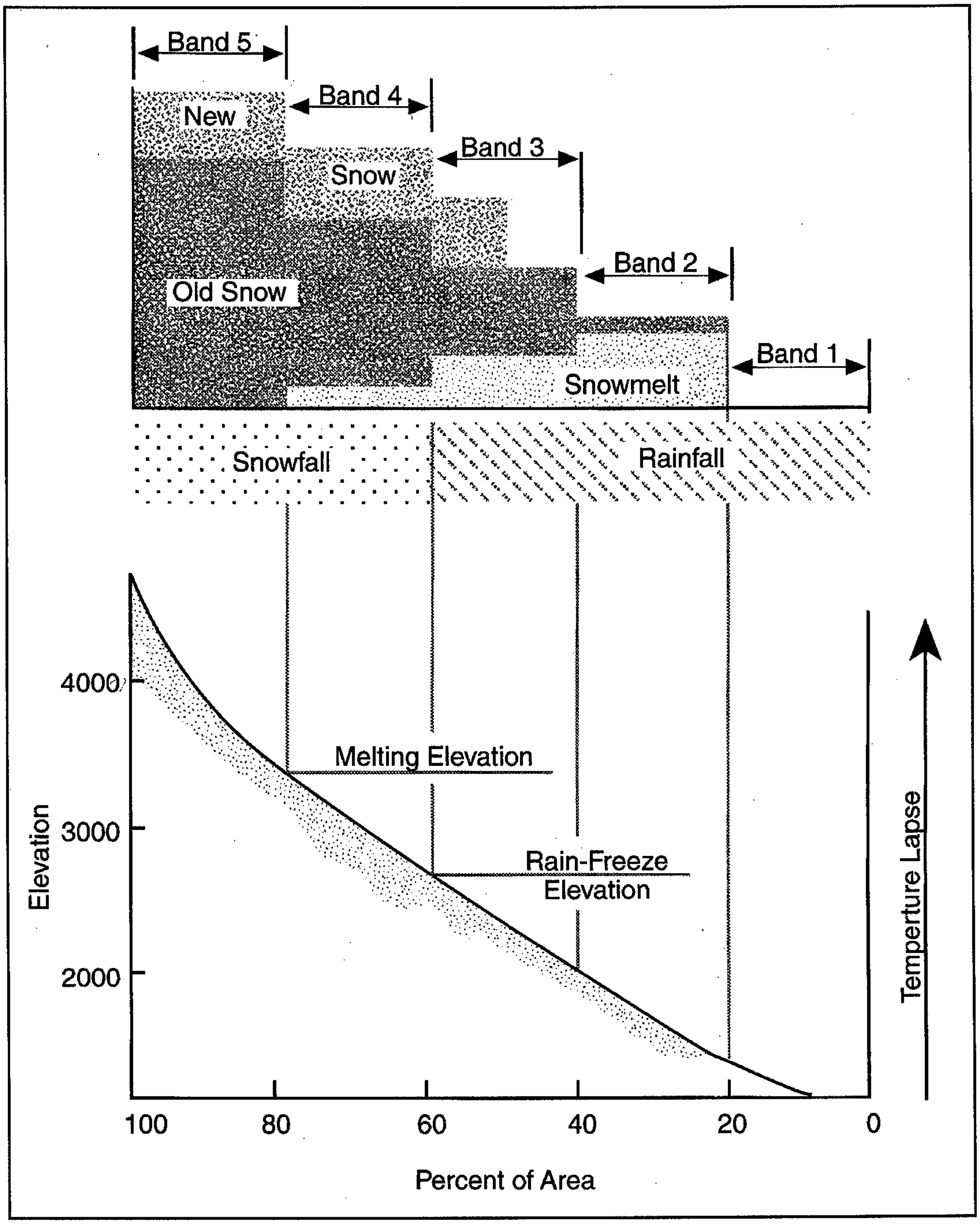

Figure 4-2. Schematic of an elevation band watershed model

(1) With the advent of digital terrain models and geographic information systems (GIS), there has been a move to define a watershed model with a fixed grid, most likely in a rectangular coordinate system. With this type of definition, such characteristics as topographic features, soil types, land-use development, and stream patterns can be specified from a GIS database. Model characteristics, including those pertaining to snow, can also be specified so that

each grid cell functions independently of the others in the simulation.

(2) Figure 4-3 is a schematic of a grid-cell basin formulation. It can be seen that the spatial, grid-cell approach can indirectly consider elevation effects. For applications in steep, mountainous terrain, the challenge for this approach is adequately defining the vertical relief. Wigmosta, Vail, and Lettenmaier (1994) employed a spatially distributed, physical model on a 2900-km^2 watershed in northwestern Montana, using a 180-m grid spacing. This requires over 220 000 cells to define the watershed.

(3) Another technique of defining a watershed is that employed by the U.S. Geological Survey and others (Leavesley et al. 1983, Kite and Kouwen 1992), where the basin is divided into relatively homogeneous HRUs based on elevation, slope, aspect, and vegetation. The Precipitation-Runoff Modeling System (PRMS) program uses this technique (see Chapter 11 and Appendix F regarding computer programs).

4-4. Design Floods—SWE Estimates from Historical Records

a. General. Certain hydrological engineering analyses require the determination of a design flood by way of applying precipitation of a specified magnitude to a rainfall-runoff model. If a snowpack is involved, the magnitude and distribution of the SWE is needed as input to the snowmelt portion of the runoff model. The SWE might best be determined by continuous simulation as described in Paragraph 4-5; however, if a continuous model is not being used, then the SWE has to be determined by an independent analysis of historical data. The SWE might either be a single basin-average value for input into a lumped model, or SWE values might be distributed into a spatial grid or elevation bands for use in a distributed melt model. The former approach, for example, would be appropriate for a relatively flat Midwest basin, while the latter method would be needed for a mountainous western basin. The values typically needed are a seasonal accumulation of winter snow, for example:

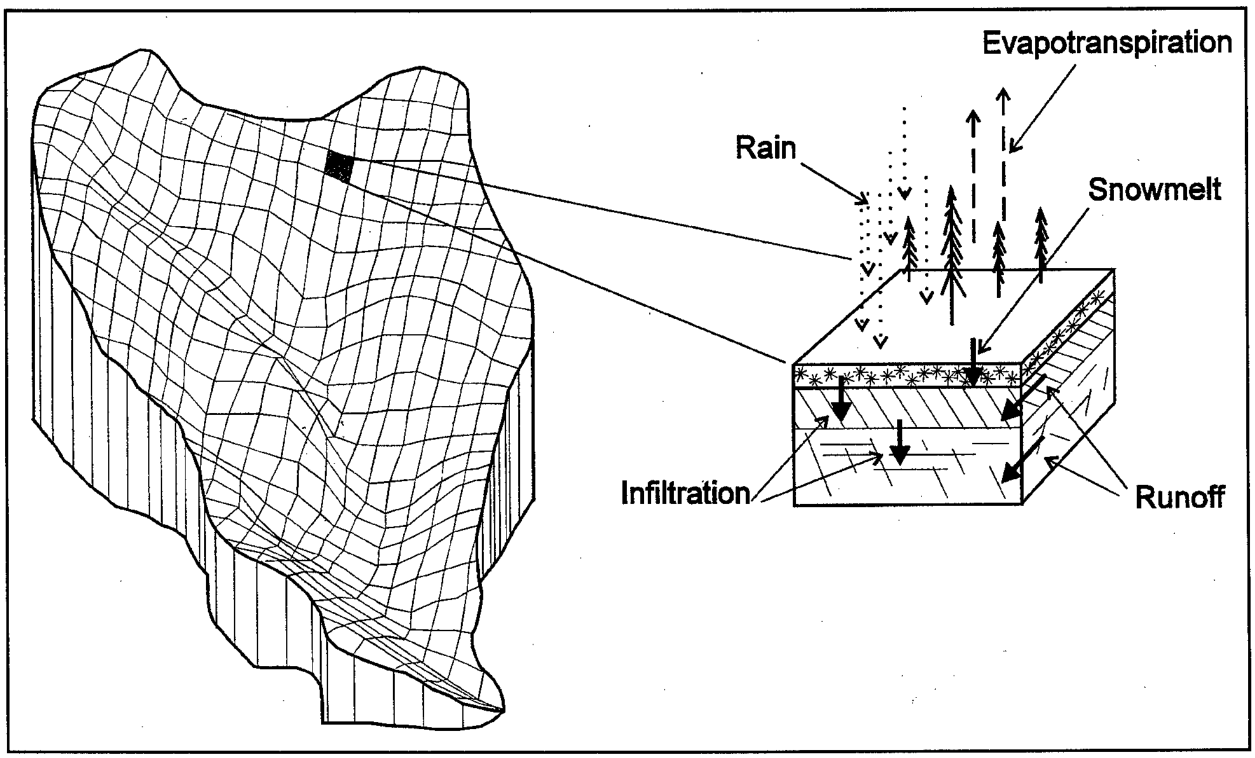

Figure 4-3. Schematic of grid cell formulation

(1) A winter (November-March) accumulation of snow as input into a spring runoff derivation.

(2) A representative midwinter accumulation (November-December) to be a factor in a December (rain-on-snow flood derivation).

b. Analysis process. The process for developing an SWE quantity is much the same as in rainfall analysis leading to input to a rainfall-runoff model. For rain analysis, the steps are as follows:

(1) Develop depth-duration-frequency curves for stations in the basin and determine the values of precipitation appropriate for the flood magnitude being analyzed.

(2) Using techniques such as the Thiesson polygon or isohyetal analysis, develop mean basin (or subbasin increment) values.

(3) Based upon historical records or design flood guidance, develop temporal distributions of the rainfall totals.

(1) For estimates of initial SWE, the first step above could involve long-term (e.g., 1-6 months) durations representing snow accumulation over all or part of a winter season. This would use available SWE records in and near the basin and would also employ precipitation data where feasible. The second step, developing areal quantities, requires more judgment and care than in rain-only cases, and most always would require an isohyetal analysis in mountainous areas. The third step above is not necessary since all that is required is an accumulated value for an initial value. Temporal distribution is determined later during snowmelt by the temperature and precipitation pattern employed as input.

(2) The difficulty in making point-to-areal SWE conversions in a mountainous winter rain-on-snow environment is illustrated in Figure 4-4. This shows the basin divided into three zones, each needing to be considered differently in the analysis. The highest parts of the basin are essentially always snow-covered in the winter, and in fact might accumulate more snow during even a relatively warm frontal passage. In this zone, the SWE determination is not particularly sensitive since there might likely be more snow present than can be melted in a 2- or 3-day rain. The lowest zone, by contrast, is essentially always snow-free, except in rare cases. It also is not critical in the analysis since any snow that would be there in some years would be shallow and assumed to be quickly melted before the peak of the flood.

(3) It is in the middle zone of Figure 4-4 that an SWE determination requires particular care. The historical records might say that in some years this is snow-free by midwinter, while in other years there is partial or complete snow cover. The analysis must determine the appropriate degree of SWE and cover associated with the given magnitude of event. Interpolation using isohyetal analysis may be difficult if, for instance, available snow gauges are located only at higher elevations, thereby not completely reflecting the conditions in the middle zone. To do a detailed determination of SWE for model input in such a situation, the best type of analysis would be continuous simulation of the period of record throughout the winter, as is described in Paragraph 4-6. For the maximum design floods, conservative estimates of the snow "wedge" could be employed as described in Chapter 10.

4-5. Forecasting Applications—SWE Estimates from Real-Time Data

Determining SWE accumulation in forecasting models theoretically employs the same process as used for design floods described above, except that the source of data is a real-time gauging network. However, given the typical uncertainties with data in a forecasting situation and the need for a quick response in making the forecast, it is quite likely that any detailed analysis will be minimal and the estimate of SWE will be relatively rough. The degree of accuracy depends heavily on the thoroughness of the real-time gauging network, and that in turn relates to the network design and the perceived need for SWE data in the forecasts. If snowmelt figures significantly in the streamflow forecasts, then the network should include strategically placed snow pillows or precipitation gauges to provide data for the model input. It would be best in such situations to have gauges in the transitory zone rather than at higher elevations where snow is always present (refer again

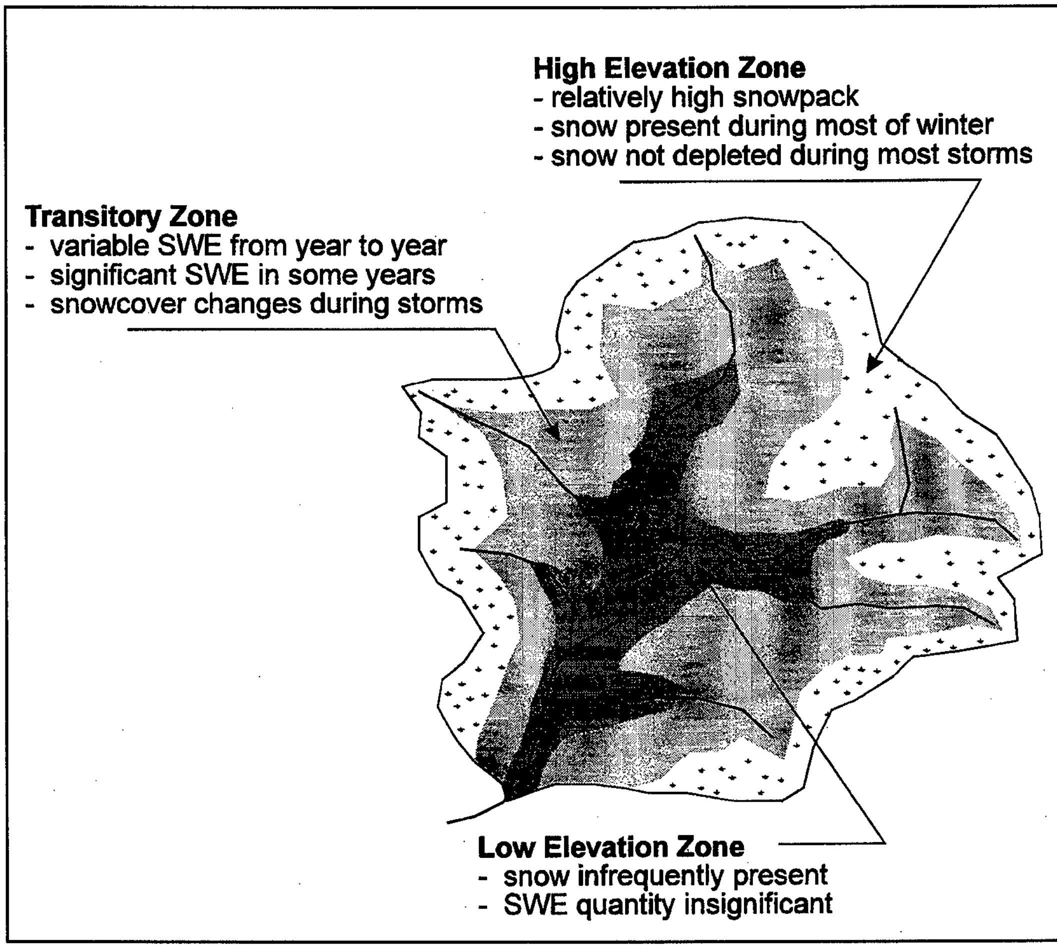

Figure 4-4. Illustration of SWE variation in a mountainous basin with rain on snow

to Figure 4-4). On the other hand, if snowmelt is a relatively small quantity compared with rainfall, the installation of snow pillows may not be warranted. Of course, only rough estimates of SWE would be possible in this case.

a. Basin-average SWE or SWE distribution can be estimated using the concept of a real-time observation acting as an index to the objective SWE variable. This requires analysis of historical data, typically using single or multiple regression. Independent variables would be the station observations available, conceivably including snow pillow, precipitation, and perhaps temperature data. The dependent variable would be basin-mean or subbasin quantity; for instance, the average SWE on a certain elevation zone in the forecast model. This technique is discussed further in conjunction with continuous model forecasting. Even if the accuracy of such relationships is relatively low, they do give a forecaster quick guidance in what may be a stressful forecast situation. In spring/summer snowmelt settings, where an extensive snow-covered area exists, the index concept can be carried to more rigorous levels by employing the advanced statistical techniques described in Chapter 9. Here, several index stations, including precipitation and SWE sensors, can be used to produce a mean basin SWE estimate for input into a snowmelt model.

b. For rough estimates of SWE where real-time SWE data are not available, the forecaster might employ SWE observations outside of the basin and manual observations of snowline elevation and snow depth from dam tenders, weather stations, ski areas,

etc. Precipitation and temperature gauge data could also be employed to keep a running estimate of snow accumulation in certain critical elevation zones—this would be a manual or spreadsheet calculation that amounts to a simple version of modeling snow accumulation with continuous simulation.

4-6. Simulation of Snow Accumulation Using Continuous Modeling

The most thorough procedure for estimating snow accumulation is to employ a continuous simulation model that operates through the winter accumulation season. The model typically uses temperature and precipitation as input and, operating on a relatively short time-step, keeps a running accounting of SWE for each of the distribution elements in the model configuration. Other phenomena that also need to be accounted for are interception and sublimation. The advantage of this approach is that the basin's SWE distribution is relatively accurately defined for the snow runoff determination involved. The disadvantage is that it requires more effort to set up and run the model and may represent "overkill" for the application involved. Figure 4-5 illustrates the steps involved in such a simulation, this case being for a snow-band model. Figure 4-6 shows the basin summary output from the SSARR model for a period of simulation during the winter. The status of each of 10 bands is shown on the right side of the output. If desired, the modeler can request a detailed listing of the computation for each of the bands.

a. For applications in hydrological engineering analysis, it is common to simulate snow accumulation and melt for a continuous period of several years, perhaps the period of record. If a long period of record is available, the statistical reliability of the SWE distribution may be relatively good. For example, in a design flood determination, the simulation results for each distribution element could be extrapolated as desired to a desired frequency level for input into a hypothetical design flood. For operational studies involving water supply and multiple-year droughts, a continuous simulation approach is almost essential if runoff modeling is required. An example of modeling for a reservoir operations study is described in Chapter 10.

b. In forecasting applications, continuous simulation can be usefully employed to obtain a distributed portrayal of SWE in the basin. It is an essential part of long-range Extended Streamflow Prediction forecasting described briefly in Chapter 10. In rain-on-snow settings, where a quick forecast response is required and snowmelt is not a key factor, the more time-consuming effort involved in running the model may limit its use in real-time in favor of the more approximate procedures described above. A continuous model could conceivably be operated as a background analyzer between forecasts, to provide an update on SWE and other variables for the forecaster, and then as an event-type model operated to produce the rain and snowmelt-runoff forecast.

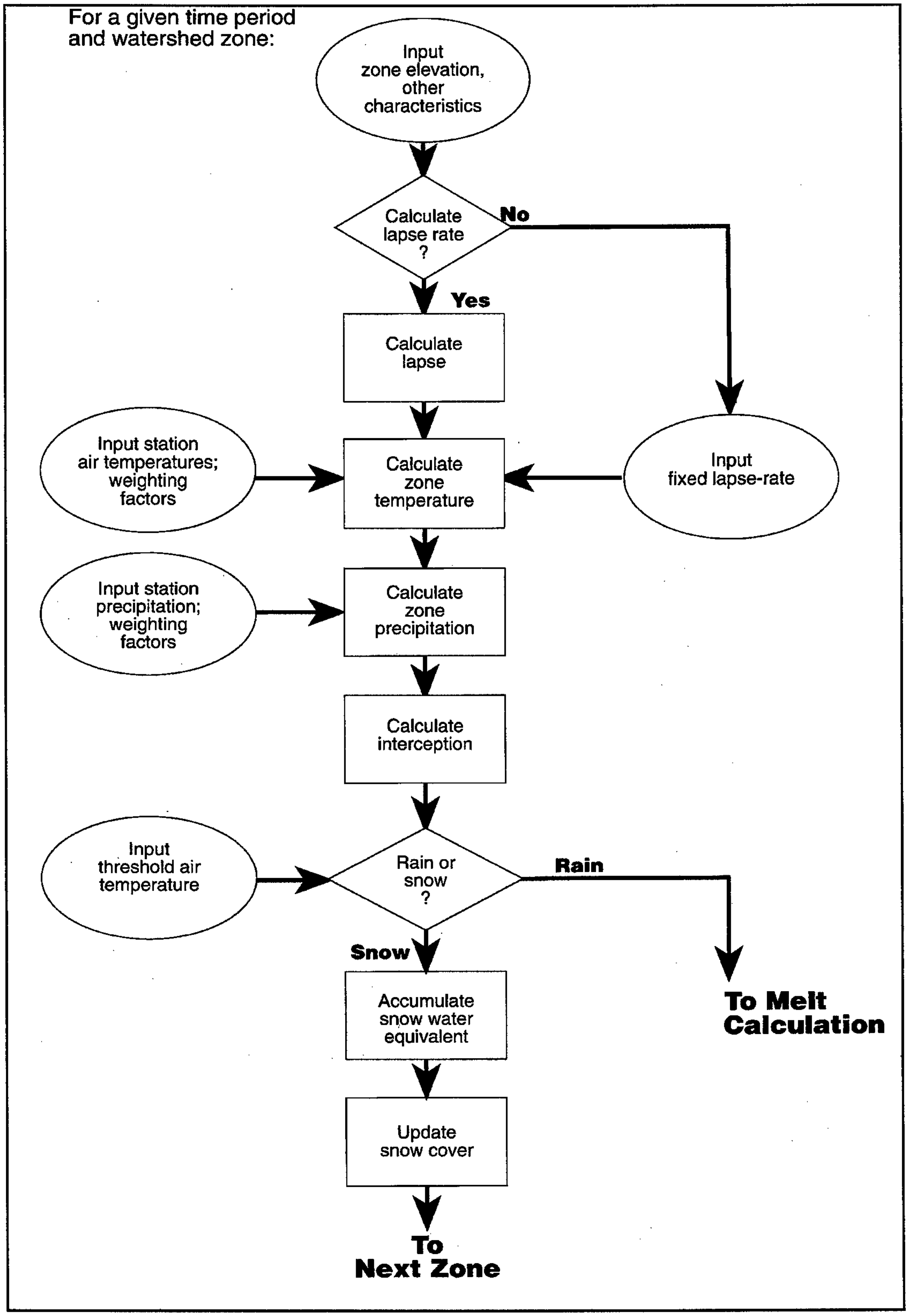

Figure 4-5. Algorithm of snow accumulation variation

SSARR SNOWBAND MODEL (METRIC) -

COMPUTED FLOW, ILLECILLEWAET R., CANADA

		AREA		BASE-TEMP			ZONES													FLAGS BY ZONE
OCT 1966		1155		0			6													
DA	HR	PCPN	INT	SNOWL	WE	LR	TA	MR	RG	ET	SMI	ROP	BFP	SURF	SUBSF	BASEF	LOWERZ	TOTAL	OBS	1.....6
1	240	0.02	0.01	2158	420	5.6	13	0.201	0.04	0.05	17	67	90	0.05	5.80	15.91	6.98	28.74	38.94	AA
2	240	0	0	2158	420	6.4	13	0.198	0.02	0.04	17	67	89	0.12	3.38	15.77	6.97	26.24	34.26	S
3	240	0	0	2158	420	6.4	12	0.195	0.01	0.04	17	67	89	0.11	2.00	15.59	6.95	24.65	28.86	S
4	240	0	0	2158	420	4.2	13	0.194	0.08	0.07	17	67	89	0.25	1.34	15.37	6.94	23.90	27.23	D
5	240	1.67	0.37	2158	420	0.8	11	0.194	1.03	0.14	17	66	88	2.73	2.81	15.34	6.92	27.80	29.52	D
6	240	2.98	0.11	2158	420	3.0	11	0.372	2.09	0.07	18	68	70	14.19	11.32	15.77	6.91	48.19	40.78	R
7	240	0.57	0.06	2158	420	6.4	16	0.191	0.35	0.07	18	71	60	18.22	19.32	16.47	6.89	60.90	55.51	DA
8	240	0.03	0.03	2158	420	6.4	11	0	0.00	0.03	18	71	65	8.74	18.17	17.05	6.88	50.84	48.99	AA
9	240	0.74	0.06	1902	420	6.4	11	0	0.21	0.03	18	71	71	3.54	14.04	17.52	6.87	41.97	33.70	AAA
10	240	0.25	0.03	1050	420	5.4	6	0	0.02	0.02	18	72	76	1.87	10.23	17.89	6.85	36.84	29.28	AAAAA
11	240	0.87	0.01	1472	420	6.4	10	0.053	0.31	0.03	18	72	82	1.48	7.52	18.19	6.84	34.03	27.02	DDCAA
12	240	0	0	1472	420	6.4	8	0.052	0.03	0.02	18	72	79	1.43	5.80	18.44	6.83	32.49	24.89	SSSS
13	240	0	0	1902	420	6.4	8	0.052	0.01	0.02	18	72	83	0.62	4.01	18.58	6.81	30.02	22.98	SSSS
14	240	0.08	0.05	1472	420	6.4	8	0	0.01	0.02	18	72	86	0.23	2.60	18.62	6.80	28.25	21.90	AAAA
15	240	2.33	0.03	1472	422	6.4	10	0.050	0.21	0.02	18	72	88	0.41	1.91	18.63	6.79	27.73	21.21	AAAA
16	240	0.31	0.02	1050	423	6.4	7	0	0.02	0.02	18	73	85	0.52	1.59	18.59	6.77	27.48	20.84	AAAAA
17	240	1.71	0.01	1050	424	6.4	8	0.049	0.06	0.02	18	73	86	0.41	1.32	18.50	6.76	26.99	20.37	ACAAA
18	240	4.38	0.01	1050	428	6.4	8	0.137	0.47	0.01	19	73	87	2.01	2.21	18.43	6.74	29.39	20.46	RCAAA
19	240	1.41	0.01	1050	429	4.2	5	0.049	0.17	0.02	19	74	84	3.56	3.79	18.40	6.73	32.47	21.01	DQAAA
20	240	1.23	0.02	1050	430	6.4	8	0.049	0.15	0.02	19	74	84	3.32	4.70	18.34	6.72	33.07	20.56	DQAAA
21	240	0.35	0.02	1050	430	6.3	6	0.050	0.03	0.02	19	74	85	2.19	4.54	18.23	6.70	31.66	19.57	AAAAA
22	240	2.84	0.01	709	433	5.4	4	0.049	0.01	0.01	19	74	86	0.88	3.49	18.06	6.69	29.13	19.29	AQAAAA
23	240	0.48	0.01	709	433	2.7	3	0.047	0.02	0.02	19	74	87	0.33	2.41	17.84	6.68	27.26	21.21	AACAAA
24	240	0.14	0.01	709	434	4.5	7	0.046	0.05	0.02	19	74	88	0.24	1.67	17.59	6.66	26.16	28.43	DDCCAA
25	240	0.68	0.02	709	434	4.5	9	0.052	0.18	0.03	19	74	88	0.57	1.51	17.34	6.65	26.06	40.07	DDDCCC
26	240	0.24	0.02	709	434	5.3	10	0.055	0.17	0.03	19	74	85	1.06	1.87	17.10	6.63	26.66	46.16	DDDCCA
27	240	0.12	0.02	709	434	6.4	10	0.052	0.10	0.02	19	75	83	1.18	2.20	16.87	6.62	26.87	40.64	DDAAAA
28	240	0.94	0.02	709	435	6.4	7	0.064	0.07	0.02	19	75	84	0.97	2.22	16.63	6.60	26.42	33.13	DAAAAA
29	240	0	0	709	435	2.6	6	0.053	0.10	0.03	19	75	86	0.82	2.11	16.37	6.59	25.88	30.44	DDDCCC
30	240	0	0	1050	434	3.4	8	0.055	0.11	0.03	19	75	85	0.82	2.04	16.11	6.58	25.55	28.53	DDDCCC
31	240	0	0	1050	434	5.0	7	0.056	0.05	0.02	19	75	85	0.72	1.89	15.85	6.56	25.02	26.47	DSSSS
		V O L U M E - C E N T I M E T E R S																		
		24.37	0.96						6.17	1.00				0.55	1.12	4.00	1.57	7.25	6.90	

EXPLANATION OF CODES

DA	Day	**SURF**	Surface flowrate, cms
HR	Hour	**SUBSF**	Subsurface flowrate, cms
PCPN	Precipitation, cm	**BASF**	Baseflow flowrate, cms
INT	Interception, cm	**LOWERZ**	Lower zone flowrate, cms
SNOWL	Elevation of snowline, meters	**TOTAL**	Total computed discharge, cms
WE	Snow water equivalent, cm	**OBS**	Observed discharge, cms
LR	Lapse rate, degrees C / 1000 m	**FLAGS**	Indicators of snow activity on each elevation band
TA	Air temperature at sea level, degrees C	**D**	Dry weather melt occurring
MR	Melt rate, cm/degrees C-day	**R**	Rain melt occurring
RG	Runoff generated, melt+precip-int-soil loss	**S**	Snow on band, no accumulation nor melt
ET	Evapotranspiration, cm/day	**A**	Snow being accumulated
SMI	Soil moisture index, cm	**L**	Dry melt restricted by band transition
ROP	Computed runoff percent	**Q**	Melt, but no RO because of liquid water deficiency
BFP	Computed baseflow percent	**C**	Melt, but no RO because of cold content

Figure 4-6. Example of computer printout during snow accumulation

Chapter 5
Snowmelt—Energy Budget Solutions

5-1. Overview

This chapter will present one of the two basic approaches to computing snowmelt, that of using energy budget equations. With this method an attempt is made to make the solution as physically based as practicable by incorporating into snowmelt equations factors such as solar radiation, wind, and long-wave radiation exchange. The second basic method, called the temperature index solution, will be covered in Chapter 6. In that more simplified approach, air temperature is assumed to be a representative index of all energy sources so that it can be used as the sole independent variable in calculating snowmelt. In Chapters 5 and 6, discussion and guidance will be presented on the appropriate usage of either of these two approaches, and Chapter 10 contains examples of applications of both methodologies.

a. Background and perspective. Researchers have, for a long time, identified the basic energy sources involved in producing snowmelt as discussed in Chapter 2. Among the earliest of these were the USACE snow investigation studies, which were aimed primarily at providing procedures for deriving maximum design floods. These studies led to the development of several generalized energy budget equations, which will be presented in this chapter, along with a summary of the technical concepts embodied in the equations. Seen from today's perspective, the USACE energy budget equations remain as viable tools that are still referenced in textbooks, handbooks, and technical papers. More recent research—see compilations by Male and Gray (1981) and Gray and Prowse (1992)—has tended to emphasize theoretical aspects of snowmelt. Even so, an empirical aspect is often present with field and laboratory experimentation being involved. The USACE equations presented in *Snow Hydrology* generally take a further step away from the theoretical by making additional assumptions, eliminating the dependence on hard-to-obtain data where possible, and combining empirical factors for simplicity. The result is that they are reasonably easy to use in engineering applications. Recent literature typically omits this step; thus, the equations remain useful as an additional bridge between the theoretical and the practical. The equations should not be used, however, without knowledge of the basic technical concepts involved; remember that they were developed from experimental data from three field sites representing specific climatic and topographical regimes.

b. Applications. As noted above, the generalized snowmelt equations were developed primarily to derive the maximum hypothetical design floods in snow regimes. That does not preclude their use in other applications, however, and in fact the equations are included in both the HEC-1 and SSARR models for general use. However, the use of meteorological variables such as solar radiation, dew point, and wind velocity generally preclude their use for real-time forecasting or perhaps for early phases of planning and engineering studies. The equations are very useful for gaining an introductory understanding of the basic principles of snowmelt and can be useful in guiding the application of the temperature index method for forecasting and analysis. Their use in developing hypothetical design floods is quite appropriate and feasible.

5-2. Basis for Equations

a. Overview. Chapter 2 describes the fundamental processes involved in the melting of snow, and Equation 2-1 expresses the basic energy balance equation appropriate for computing snowmelt runoff. There are six external sources of heat energy represented in that equation, and these must be accounted for one way or another in developing applied snowmelt equations. The following discussion will briefly summarize the theoretical principles associated with each of these components, following up from the general description in Chapter 2, then describe the assumptions reflected in the adapted relationships that make up the generalized equations presented in Paragraphs 5-3 and 5-4. The basic source of backup information is *Snow Hydrology* (USACE 1956), unless otherwise noted. For a background on some of the basic physics principles involved, see Appendix C. Appendix D contains background on basic meteorological relationships pertaining to snow hydrology, including some pertinent charts taken from *Snow Hydrology.*

b. Units. The equations in this chapter will be presented in both SI and English units. For the discussion on the sources of the generalized energy budget equations, the SI convention will be followed as much as possible, as was done in Chapter 2. If the reader refers to modern textbooks on physics and meteorology on this subject, the SI convention would be used exclusively. However, once the discussion involves the experimental relationships that were developed in the 1950s, current U.S. practice (English units) will be followed. The generalized equations presented in Paragraphs 5-3 and 5-4 will also use the U.S. convention, since that is the current practice here. Alternative forms of these equations in SI units are given in Appendix E. A second problematic area regarding unit conventions is how heat and radiation energy are treated. In the investigations described in *Snow Hydrology*, the heat quantity calorie was used, along with the radiation term langleys (calories/square centimeter). This convention has now been replaced by the use of joules, where 1 gram-calorie = 4.186 joules. Radiation flux is currently reported in several ways, as discussed in Appendix D. A conversion table is contained in Appendix C to assist in dealing with a somewhat confusing mixture of units.

c. Shortwave radiation melt. The applied equation component for shortwave radiation melt is taken directly from the theoretical equation for net radiation energy input at a point, Equation 2-4, combined with the general formula for snowmelt, Equation 2-2.

Thus

$$M_{sw(mm)} = \frac{1000(1-a)I_i}{334.9\,\rho B} \tag{5-1}$$

where

M_{sw} = daily shortwave radiation snowmelt, mm

a = snow albedo

I_i = daily incident solar radiation, kJ/(m^2 day)

ρ = density of water, 1000 kg/m^3

B = thermal quality of the snow

334.9 = latent heat of fusion of ice, kJ/kg

When the thermal quality is assumed to be 0.97 as discussed in Chapter 2, this equation reduces to

$$M_{sw(mm)} = 0.00308 I_i (1-a) \tag{5-2}$$

The alternative equation, when melt is expressed in inches and solar radiation is expressed in langleys, is obtained by employing the equivalent of Equation 5-1:

$$M_{sw(mm)} = \frac{(1-a)I_i}{(2.54)(80)B} \tag{5-3}$$

where

M_{sw} = daily shortwave radiation snowmelt, in.

I_i = daily incident solar radiation in langleys, cal/cm^2

2.54 = converts centimeters to inches

80 = latent heat of fusion of ice, cal/cm^3

This becomes

$$M_{sw} = 0.00508 I_i (1-a) \tag{5-4}$$

which becomes part of the generalized equation for melt (inches) in an open area presented in Paragraph 5-4.

d. Long-wave radiation melt. Long-wave radiation melt equations must consider, first, the radiation to the atmosphere from the snow surface, resulting in a net energy loss on clear days, and, second, the incoming (back) radiation emitted by the Earth's atmosphere, cloud cover, and forest canopy. Since the snow surface is nearly a perfect blackbody source of radiation, with a maximum temperature of 0 °C, long-wave radiation from the snow surface can be

expressed as a constant employing the Stefan-Boltzmann equation. From Equation 2-5, using an emissivity of 0.99, this has been computed to be 0.315 kJ/(m^2 s). Using the older units of calories and langleys, the Stefan-Boltzmann coefficient is 8.26×10^{-10} ly/(min K^4), and the equation produces a long-wave radiation flux of 0.459 ly/min. This value is used in the generalized equation development that follows. It assumes an emissivitity of 1.0. Gray and Prowse (1992) note that emissivities can vary from 0.97 for dirty snow to 0.99 for clean snow.

(1) Back-radiation is a complex phenomenon involving factors such as the temperature of the cloud cover and tree canopy and the distribution of water vapor and temperature in the atmosphere. For that reason, experimental data and simplifying assumptions are used to develop relationships to express this. For back-radiation over snow under clear skies, the snow investigations experiments showed that a simple air temperature function can adequately express downward long-wave radiation because of the restricted range in vapor pressure normally experienced in these conditions. This equation is

$$Q_b = 0.76 \sigma T_a^4 \tag{5-5}$$

where

Q_b = long-wave radiation, ly/min

σ = Stefan-Boltzmann constant, ly/(min K^4)

T = air temperature, K

The net exchange by long-wave radiation is then:

$$Q_n = 0.76 \sigma T^4 - 0.459 \text{ (ly/min)} \tag{5-6}$$

(2) When clouds or forest cover are present, the back-radiation may be computed assuming that either is emitting radiation as a blackbody. Thus, the net long-wave radiation is computed by

$$Q_n = \sigma T^4 - 0.459 \quad \text{(ly/min)} \tag{5-7}$$

where T, the free air temperature, is assumed to approximate the temperature of the forest cover under surface or that of a low-elevation cloud base.

(3) The snowmelt resulting from long-wave radiation exchange is computed by combining Equations 5-6 and 5-7 with the general equation for melt, Equation 2-2. The resulting functions are nonlinear relationships between temperature (K) and long-wave radiation. In the snow investigation studies, these were simplified by fitting linear approximations and shifting to the Fahrenheit temperature scale. This is illustrated in Figure 5-1. The resulting equations for long-wave radiation melt are as follows.

(a) For melt under clear skies:

$$M_l = 0.0212 (T_a - 32) - 0.84 \tag{5-8}$$

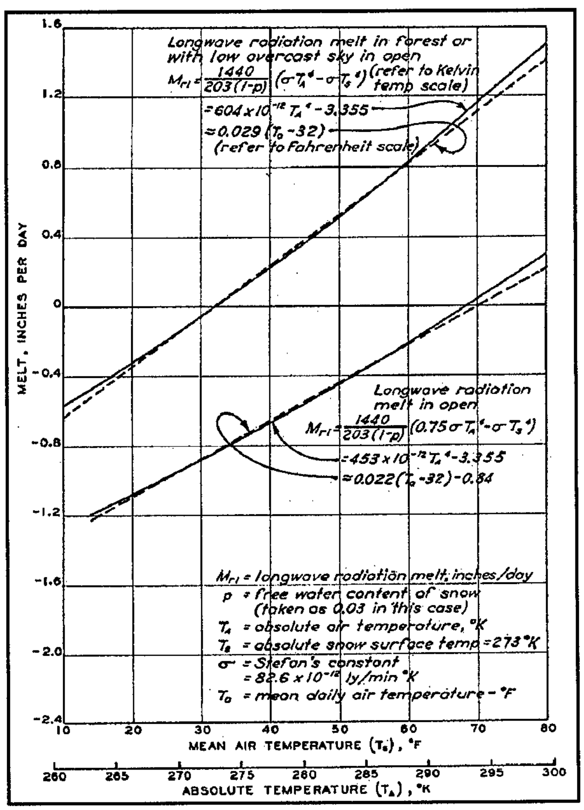

Figure 5-1. Linear adaptation of long-wave radiation functions (Figure 1, Plate 6-2, *Snow Hydrology)*

where M_l equals long-wave radiation melt (inches/day).

(b) For melt under a forest canopy or under a low cloud cover, as would be experienced during rain on snow,

$$M_l = 0.029(T_a - 32) \qquad (5\text{-}9)$$

e. Convection (sensible heat) melt. Equation 2-6 is a general equation widely used to express the convective heat transfer between the air and snow surface. It represents a simplification of a complex physical process involving turbulent exchange taking place in the atmosphere 2 to 3 m above the snow surface. The key to this equation is the bulk transfer coefficient D_x, which has to be determined experimentally. As pointed out in Male and Gray (1981), there is a wide range of variation in the coefficient reported by researchers, so it is fortunate that the magnitude of this component of snowmelt is relatively small. This reference compares values from various sources including the snow investigations laboratories.

(1) The bulk exchange coefficient arrived at in the snow investigations program was based upon observations taken at the Central Sierra Snow Laboratory. Two other factors are also introduced to express the density of the atmosphere and to account for differences in the heights at which temperature and wind speed are measured. The resulting equation is similar in form to Equation 2-6 but is expressed directly in terms of snowmelt by applying the basic equation for snowmelt at a point (Equation 2-2):

$$M_c = 0.00629\left(\frac{p}{p_o}\right)(z_a z_b)^{-1/6}(T_a - T_s)\,v_b \qquad (5\text{-}10)$$

where

M_c = convection melt, in./day

p, p_o = atmospheric pressures at location and at sea level, respectively

T_a = air temperature, °F

T_s = snow surface temperature, generally 0 °C (32 °F)

v_b = wind velocity, miles/hour

z_a,z_b = height of temperature and wind velocity measurement, ft

(2) For snow hydrology applications, Equation 5-10 was further simplified by assuming measurement heights of 3 and 15.2 m (10 and 50 ft) for air temperature and wind velocity, and by assuming a constant value of 0.8 for the atmospheric pressure ratio. This value would be considered appropriate for mountainous regions, with the range being 1.0 at sea level to 0.7 at a 3048-m (10,000-ft) elevation. With these assumptions, Equation 5-10 becomes

$$M_c = 0.00179 v_b (T_a - 32) \qquad (5\text{-}11)$$

f. Condensation (latent heat) melt. The equation for computing condensation melt is similar in form to that for convection melt. Equation 2-7 defines the basic form, and the bulk transfer coefficient is determined from field measurements. The snow investigation studies led to the following equation based upon experimental analysis at the Central Sierra Snow Laboratory:

$$M_e = 0.054(z_a z_b)^{-1/6}(e_a - e_s)v_b \qquad (5\text{-}12)$$

where

M_e = condensation melt, in./day

z_a,z_b = measurement heights, feet above snow surface for air vapor pressure and wind speed, respectively

e_a = vapor pressure of the air, in.

e_s = vapor pressure of the snow surface, mb

v_b = wind velocity, miles/hr

(1) Figure 5-2 is a plot of this equation, assuming a vapor pressure difference at 0.3 m (1 ft) above the

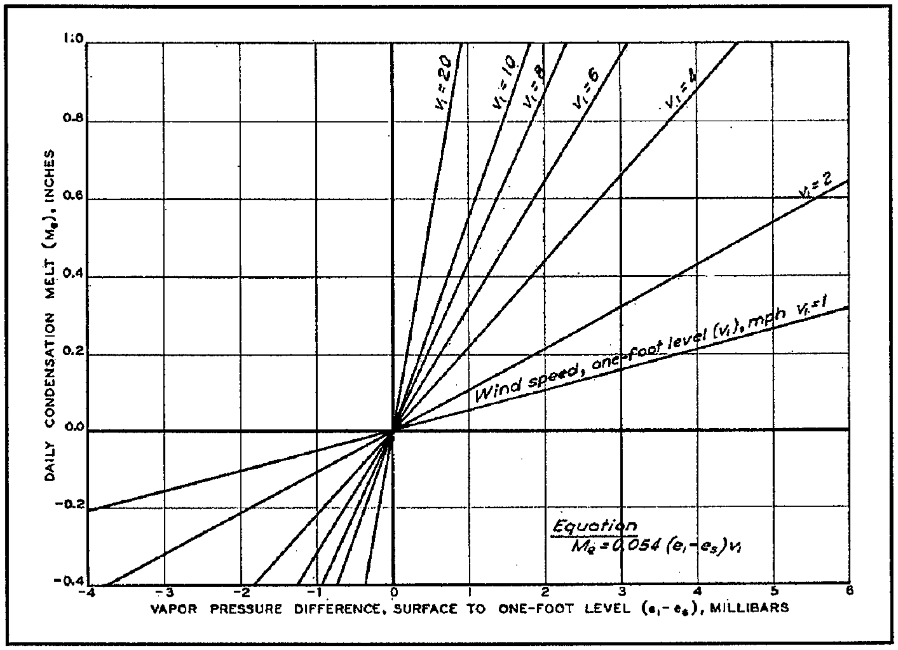

Figure 5-2. Daily condensation melt versus pressure gradient (Figure 2, Plate 5-5, *Snow Hydrology*

snow surface. Note the negative range of the function, indicating evaporation from the snow surface.

(2) Equation 5-12 can be simplified by assuming standardized measurement heights of 3 and 15.2 m (10 and 50 ft) above the snow surface, as was done with Equation 5-10. The other simplifying step is to replace vapor pressure with a variable that can be more practically measured and applied. A useful relationship exists between vapor pressure and dew-point temperature as shown in Figure 5-3. For the range of the variables normally encountered in practice, a linear approximation can be fitted:

$$e = 6.11 + 0.339(T_d - 32) \qquad (5\text{-}13)$$

where

e = vapor pressure, mb

T_d = dew-point temperature, ° F

6.11 = saturation vapor pressure, mb

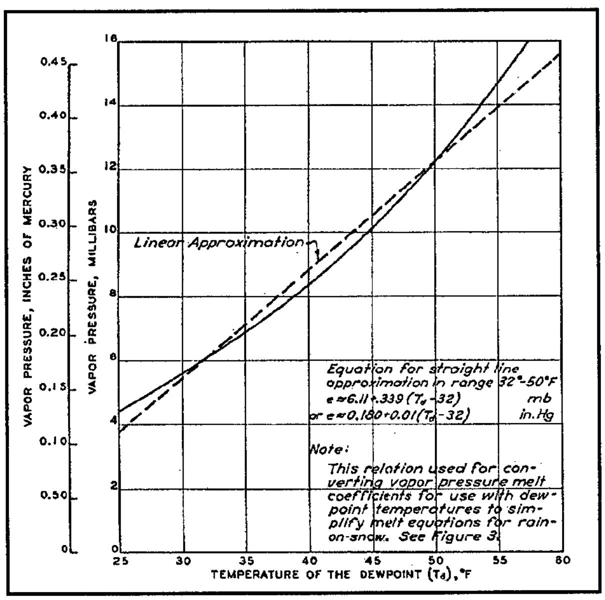

Figure 5-3. Experimental relationship between vapor pressure and dew point (Figure 5, Plate 6-2, *Snow Hydrology)*

(3) Combining Equations 5-12 and 5-13 and assuming the vapor pressure of the snow surface to be 6.11 mb, results in the simplified equation for condensation melt

$$M_e = 0.0065(T_d - 32)v_b \tag{5-14}$$

g. Combined convection-condensation equation. Since the equations for convection and condensation melt share some of the same variables, they are often shown in a combined form. Adding together Equations 5-11 and 5-14, the equation for combined daily snowmelt attributable to convection and condensation can be written:

$$M_{ce} = 0.0084v[0.22(T_a - 32) + 0.78(T_d - 32)] \tag{5-15}$$

where M_{ce} equals combined convection-condensation melt (inches per day).

In heavily forested areas where wind effects can be considered negligible, an alternative to Equation 5-15 for combined convection-condensation melt was determined experimentally:

$$M_{ce} = 0.045(T_a - 32) \tag{5-16}$$

h. Rain melt. Equation 2-9 is the basic formula expressing the heat energy given up when rainwater is cooled to the temperature of the snowpack, assuming the snowpack temperature is 0 °C. Using the following values for the coefficients,

$\rho = 1000$ kg/m^3

$C_p = 4.20$ kJ/(kg °C)

$T_s = 0$ °C

and applying Equation 2-2, this equation becomes

$$M_{r(mm)} = 0.0125T_rP_r \tag{5-17}$$

where

M_r = daily snowmelt from heat supplied by rain, mm

T_r = temperature of rain, °C

P_r = daily rainfall, mm

The alternative to Equation 5-17 for English units is

$$M_r = 0.007(T_p - 32)P_r \tag{5-18}$$

where

M_r = daily snowmelt from heat supplied by rain, in.

T_r = temperature of rain, °F

P_r = daily rainfall, in.

i. Ground melt. The final source of energy for snowmelt is heat conducted from the ground. Once a snowpack becomes deep enough to insulate the ground from subfreezing air, an upward flux of heat can act to melt snow at the bottom of the snowpack. Although the rate of heat exchange is small, it can act continuously throughout a winter. As discussed in Chapter 2, a constant value is typically assumed for this component. Field experiments reported in *Snow Hydrology* and by Male and Gray (1981) estimate melt rates of 0.025 to 0.076 cm/day (0.01 to 0.03 in./day) ascribable to ground heat.

5-3. Generalized Equations, Rain-on-Snow Situations

a. Overview. For practical engineering use, the equations for snowmelt presented above can be combined into several generalized equations designated for specific meteorological and forest-cover conditions. Often the equations can be further simplified when the application is limited as specified. Also

covered in this paragraph and in Paragraph 5-4 is the need to consider the equation being applied to a basin area rather than at a point, which is the basis for its derivation. This is accomplished by introducing constants representing the mean basin exposure to solar radiation or wind. This paragraph will present equations for use in rain-on-snow conditions, with varying degrees of forest cover. Paragraph 5-4 will introduce similar equations for rain-free applications.

b. Classification of forest density. The generalized equations presented below and in Paragraph 5-4 have been adopted to varying degrees of forest cover in the basin. Table 5-1 is a general guideline to follow in selecting the appropriate equation.

Table 5-1
Classification of Forest Density

Descriptive Category	Mean Canopy Cover, %
Heavily forested	>80
Forested	60-80
Partly forested	10-60
Open	<10

c. Basin wind exposure coefficient, k. For convection-condensation melt in basins, it is necessary to introduce a basin constant, k, that represents the mean exposure of the basin, or a segment of it, to wind, considering topographic and forest effects. For unforested plains, k would be 1, but for forested areas, the value may be as low as 0.3, depending upon the density of the forest stands. This factor can be estimated from topographic maps and aerial photographs but is best confirmed through model calibration.

d. Generalized equations. Snowmelt calculation in rain-on-snow settings is the simplest application of energy budget equations since solar radiation is minimal and the atmosphere can be assumed saturated, thereby simplifying the computation of convection and condensation melt. Two equations have been developed for rain-on-snow situations. The assumptions reflected in these equations follow and are summarized in Table 5-2 (Paragraph 5-5). Appendix E contains versions of these equations in SI units.

For open or partly forested basin areas,

$$M = (0.029 + 0.0084kv + 0.007P_r)(T_a\text{-}32) + 0.09 \qquad (5\text{-}19)$$

For heavily forested areas,

$$M = (0.074 + 0.007P_r)(T_a\text{-}32) + 0.05 \qquad (5\text{-}20)$$

where

M = snowmelt, in./day

k = basin wind coefficient

v = wind velocity, miles/hr

P_r = rate of precipitation, in./day

T_a = temperature of saturated air, at the 3-m (10-ft) level, °F

e. Open-partly forested basin equation. This equation is based upon simplified equations introduced in Paragraph 5-3. Shortwave radiation has been assumed constant at 0.127 cm/day (0.05 in./day), and ground melt is assumed to be 0.05 cm/day (0.02 in./day). Long-wave radiation uses Equation 5-9. The atmosphere is assumed to be saturated for these conditions, enabling the equating of dew-point temperature in Equation 5-15 to air temperature. This equation then becomes M_{ce} = 0.0084 (T_a - 32). Rain melt is computed with Equation 5-18, assuming that the rainwater temperature is equal to air temperature.

f. Heavily forested basin equation. Because of the dense forest cover, wind is assumed to be negligible in the convection-condensation equation. This permits using the alternative, Equation 5-16. A slight reduction is made in the assumed shortwave radiation to 0.076 cm/day (0.03 in/day).

g. Measurement height adjustment. As discussed in Paragraph 5-3, the convection and condensation equations reflect a simplifying assumption to the more

Table 5-2
Summary of Generalized Snowmelt Equations, Rain-on-Snow Situations

Equation	$M = (0.074+0.007P_r)(T_a-32)+0.05$	$M = (0.029+0.0084kv+0.007P_r)(T_a-32)+0.09$
Forest-Cover Application	Heavily forested (>80% cover)	Open to partly forested (10-80% cover)
Shortwave Radiation	• Very minor contribution • Assumed constant: 0.076 cm/day (0.03 in./day)	• Minor contribution • Assumed constant: 0.05 cm/day (0.02 in./day)
Long-wave Radiation	• Relatively important • Estimated as function of air temp.—factor is 0.029 in 0.074 coefficient • See Para. 5-2d; Equation 5-9 • Ref *Snow Hydrology* (SH), Ch. 6; Plate 6-2	• Relatively important • Estimated as function of air temp. (0.029) • See Para. 5-2d; Equation 5-9 • Ref. SH, Ch. 6, Plate 6-2
Convection-Condensation	• Relatively important melt component • Wind not a factor because of forest • Estimated as a function of air temp—factor is 0.045 in 0.074 coefficient • Conv. melt factor is $0.010T'_a$ • Cond. melt factor is $0.035T'_a$ • See Equation 5-16 • Ref SH, p. 231, Plate 6-2/Fig. 3	• Wind is an important factor • Estimated as a function of wind and air temp—coefficient = 0.0084 • Conv. melt factor = $0.0018T'_a v$ • Cond. melt factor = $0.0066T'_a v$ • Need to estimate k - basin exposure to wind. Varies 0.3 to 1.0 • Dew-point temp. assumed equal to air temp. (100% relative humidity) • See Equation 5-15 • Ref SH, Ch. 6, p. 231 • Ref Male and Gray (1981), pp. 385-393
Rain Melt	• Relatively small factor ($0.007P_rT'_a$) • Based upon heat content in rain, assuming rain temp. = air temp. • See Equation 5-18 • Ref SH, pp. 180, 230	• Relatively small factor ($0.007P_rT'_a$) • Based upon heat content in rain, assuming rain temp. = air temp • See Equation 5-18 • Ref SH, pp. 180, 230
Ground Melt	• Assumed constant: 0.05 cm/day (0.02 in./day)	• Assumed constant: 0.05 cm/day (0.02 in./day)

basic turbulent transfer equations that temperature and dew point and wind speed measurements are at 3 and 15.2 m (10 and 50 ft) above the snow surface, respectively. This assumption makes use of the relationship that defines the temperature and vapor pressure profiles as varying in height according to a 1/6 power function (*Snow Hydrology*, Chapter 5, USACE 1956). If measurements are made at heights other than the assumed 3 and 15.2 m (10 and 50 ft), the following adjustment factors can be used:

$$\text{Air temperature:}\quad CF_a = 1.47Z_a^{1/6} \tag{5-21}$$

$$\text{Wind velocity:}\quad CF_b = 1.92Z_b^{1/6} \tag{5-22}$$

where Z_a and Z_b are the height of the measurement above the snow surface in feet.

5-4. Generalized Equations, Rain-Free Situations

a. Overview. In rain-free settings, the calculation of snowmelt with energy budget equations must include solar radiation as a variable (unless there is heavy forest cover) in addition to the components considered in rain-on-snow situations. This introduces additional variables, such as albedo and cloud cover, as well as new factors that are needed to convert equations for melt at a point to a basin-mean relationship. Also, a saturated air mass can no longer be assumed, thus requiring use of dew point as a variable. These variables and coefficients will be described in this chapter, and the generalized equations will be presented along with a summary of the assumptions reflected in each equation. A tabular summary (Table 5-3) is presented.

Table 5-3
Summary of Generalized Snowmelt Equations, Rain-Free Situations

Equation	$M=0.074(0.53T'_a+0.47T'_d)$	$M=k(0.0084v)(0.22T'_a+0.78T'_d)+0.029T'_a$
Forest Cover Application	Heavily forested (>80% cover)	Forested (60-80% cover)
Shortwave Radiation Melt; Ground Melt	• Relatively unimportant; assumed compensated for by evapotranspiration	• Relatively unimportant; assumed compensated for by evapotranspiration
Long-Wave Radiation Melt	• Relatively important • Estimated as function of air temp.—factor is $0.029T_a$ • See Para. 5-2d, Equation 5-9 • Ref SH, Plate 6-2	• Relatively important • Estimated as function of air temp.—factor is $0.029T_a$ • See Para. 5-2d, Equation 5-9 • Ref SH, Plate 6-2
Convection-Condensation Melt	• Relatively important • Wind not a factor because of forest cover • Conv. estimated as a function of air temp.—factor is $0.011T_a$ • Cond. estimated as a function of dew-point temp.—factor is $0.035T_d$ • Ref SH, Plate 6-2/Fig. 3	• Relatively important • Wind is an important factor • Conv. estimated as a function of air temp. and wind—factor is $0.00187T'_a v$ • Cond. estimated as a function of dew-point temp. and wind—factor is $0.0066T'_d v$ • Need to estimate k - basin exposure to wind. Varies 0.3 to 1.0 • See Para. 5-2e,f; Equations 5-11, 5-13 • Ref SH, Plate 6-2/Fig. 3 • Ref Male and Gray (1981), pp. 385-393
Equation	$M=k'(1-F)(0.0040I_i)(1-a)$ $+k(0.0084v)(0.22T'_a+0.78T'_d)$ $+F(0.029T'_a)$	$M=k'(0.00508I_a)(1-a)$ $+(1-N)(0.0212T'_a-0.84)$ $+N(0.029)T'_c$ $+k(0.0084v)(0.22T'_a+0.78T'_d)$
Forest Cover Application	Partly forested (10-60%)	Open (<10%)
Shortwave Radiation Melt	• Important factor • Function of solar insolation and albedo for unforested portions of the basin • Need estimate of k' factor (see Para. 5-4d) • Long-wave loss for open areas reflected in the shortwave coefficient, 0.004 • See Para. 5-4c re: albedo • See Para. 5-4e re: forest-cover factor, F • See Para. 5-4h • Ref SH, pp. 212-214	• Important factor • Function of solar insolation and albedo • Uses theoretical melt equation (see Equation 5-4) • Need estimate of k' factor (see Para. 5-4d) • See Para. 5-2c • Ref SH, pp. 212
Long-Wave Radiation Melt	• Relatively important factor • For forested area: function of air temp.—factor is $0.029T'_a$ • For unforested area: computed indirectly by reducing SW melt factor • See Para. 5-4e re: forest-cover factor, F • See Para. 5-2d, Equation 5-9, Para. 5-4h • Ref SH, Plate 6-2	• Important factor—loss in clear areas • Computed directly for cloud-free areas—factor is $(0.0212T'_a-0.84)$ • See Para. 5-2d, Equation 5-8 • Ref SH, Plate 6-2/Fig. 1
Convection-Condensation Melt	• Less important compared with SW melt • Computed as in forested area equation	• Less important compared with SW melt • Computed as in forested area equation

b. Solar radiation. This variable, discussed in Paragraphs 2-2 and 5-3 and Appendix D, needs to be specified as input unless there is heavy forest cover. The following two basic approaches are used in preparing solar-radiation input.

(1) Observations of solar radiation are made at first-order National Weather Service stations in the United States. These data are available from regional and national NWS archives. The data are reported as insolation (shortwave solar radiation on a horizontal surface). Since there are relatively few stations making these observations, it is unlikely that historical observations would be used directly as model input (for model calibration, for instance); however, such data could be used to estimate a historical time series or to help construct a hypothetical time series for a design flood derivation.

(2) Equations, charts, and nomographs have been developed that can be used to construct hypothetical time series of daily solar radiation or as the basis for estimating maximum theoretical insolation for historical conditions. These generally involve a theoretical insolation quantity that is based upon latitude and time of year, then corrected for transmittivity through the atmosphere. Reference is made to Appendix D, which contains a chart that could be used for this, and to Male and Gray (1981). It is necessary to establish a reasonable relative magnitude for solar radiation that is consistent with the engineering application involved, e.g., a maximized sequence in the case of a probable maximum flood (PMF) derivation. The key variable affecting the quantity of solar radiation is cloud cover, once the location and time of year are established. The appropriate amount of cloud cover could be estimated by referring to historical records of sunshine duration, diurnal temperature, cloud cover, etc. Figure 5-4 is a plot, derived from Figure D-8, showing the effect of cloud cover on insolation, given a known theoretical solar radiation amount based upon latitude and time of year. An example of solar radiation sequence developed for a PMF derivation is described in Chapter 10.

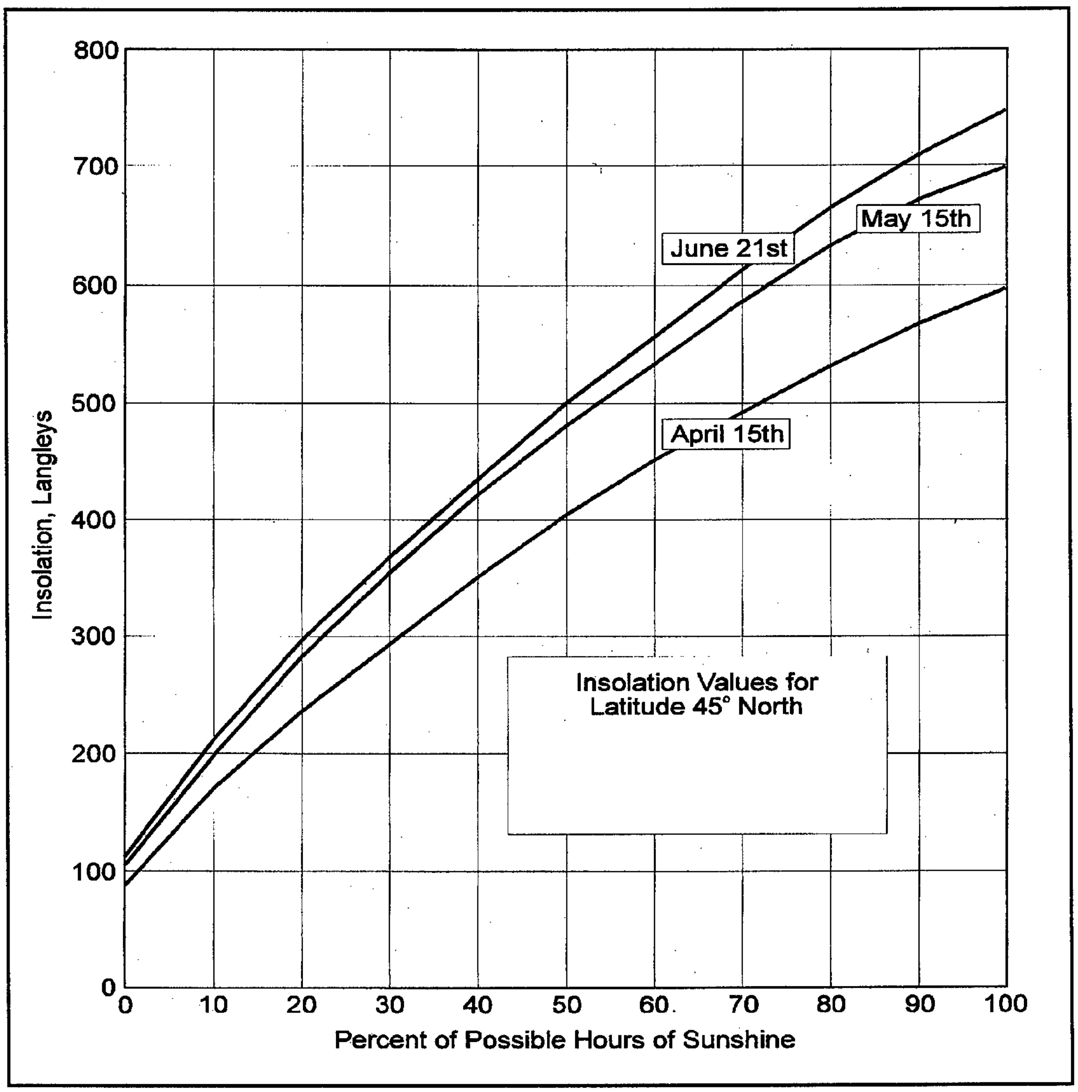

Figure 5-4. Effect of cloud cover on solar (shortwave) radiation

c. Snow surface albedo. Since there are no regular observations of snow surface albedo, this variable must be estimated on the basis of relationships established in laboratory experiments. Figure 5-5 shows a typical variation of snow surface albedo with time, for both the accumulation and melt seasons. This illustrates the general phenomenon involved, that albedo decreases as the snowpack ripens. In computer simulations this can be expressed as a decay function in the form (Laramie and Schaake 1972):

$$a = e(f)^{N^g} \qquad (5\text{-}23)$$

where

a = snow surface albedo

N = number of elapsed days

e, f, g = experimental coefficients

d. Basin shortwave melt coefficient, k'. Measurements of solar radiation are generally expressed in terms of amounts on a horizontal surface. For basins whose exposure is predominately north- or south-facing, a basin shortwave melt coefficient must be introduced in the melt equation. Reference is made to Figure D-6, showing the effect on incident solar radiation of a 25° slope at latitude 46°30′N. In general, averaged over a basin, the slope effect would not be as extreme as the particular example shown in this figure. The value of k' would be 1.0 for a basin that is essentially horizontal or whose north and south slopes are areally balanced. The value of k' usually would fall within the limits of 0.9 and 1.1 during the spring.

e. Effective forest canopy cover, F. For partly forested basins, it is necessary to estimate the effective forest canopy cover F, which is applied to determine shortwave and long-wave radiation snowmelt. The coefficient F represents the average proportion of the basin shaded by the forest from direct solar radiation, expressed as a decimal fraction. Determination of F must be based upon a partly subjective estimate of the forest characteristics, considering density and spacing of forest stands, latitudinal, and diurnal effects of the forest upon shading, and general knowledge gained from personal observation or remote sensing

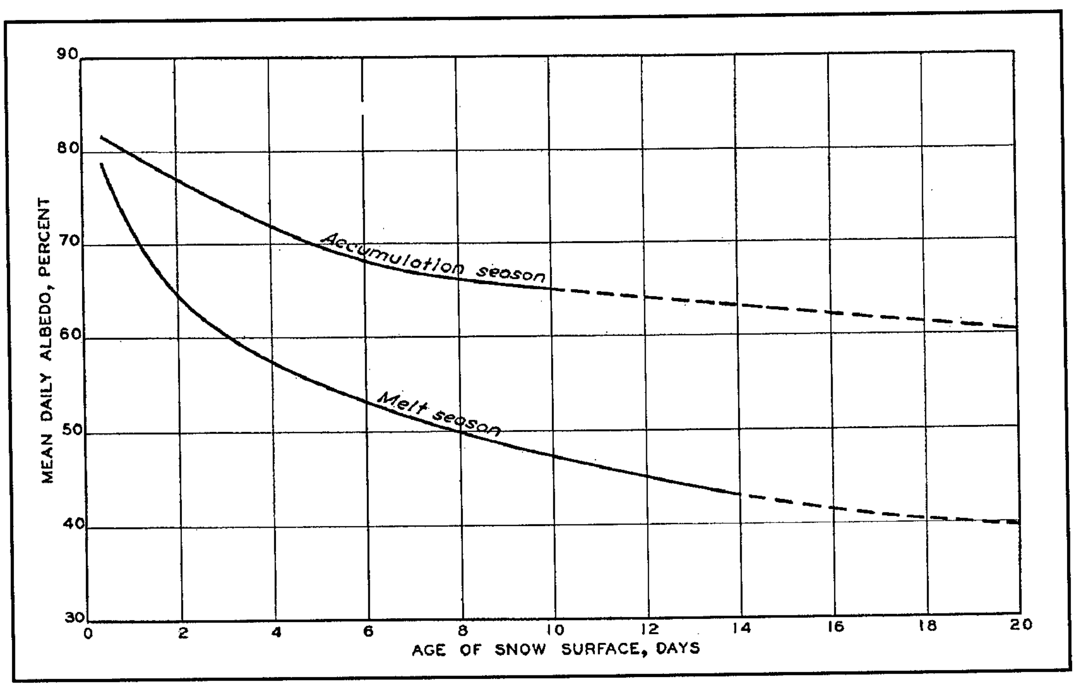

Figure 5-5. Albedo versus age of snow (Figure 4, Plate 5-2, *Snow Hydrology*)

photography. In general, the value of F is somewhat greater than the theoretical cover taken as the horizontal projection of the forest canopy (see Figure D-7).

f. Generalized equations. Four equations have been developed for the four categories of forest cover presented in Table 5-1.

(1) For open areas

$$M = k'(0.00508 I_i)(1-a) + (1-N)(0.0212\, T'_a - 0.84) + N(0.029 T'_c) + k(0.0084v)(0.22\, T'_a + 0.78 T'_d) \qquad (5\text{-}24)$$

(2) For partly forested areas

$$M = k'(1-F)(0.0040 I_i)(1-a) + k(0.0084)(0.22 T'_a + 0.78 T'_d) + F(0.029 T'_a) \qquad (5\text{-}25)$$

(3) For forested areas

$$M = k'(0.0084v)(0.22 T'_a + 0.78 T'_d) + F(0.029 T'_a) \qquad (5\text{-}26)$$

(4) For heavily forested areas

$$M = 0.074(0.53 T'_a + 0.47 T'_d) \qquad (5\text{-}27)$$

where

M = snowmelt rate, in./day

T'_a = difference between the air temperature measured at 3 m (10 ft) and the snow surface temperature, °F

T'_d = difference between the dew-point temperature measured at 3 m (10 ft) and the snow surface temperature, °F

v = wind speed at 15.2 m (50 ft) above the snow surface, miles/hr

I_i = insolation (solar radiation on a horizontal surface, langleys

a = average snow surface albedo, decimal fraction

k' = basin shortwave radiation melt factor

F = average basin forest-canopy cover shading of the area from solar radiation, decimal fraction

T'_c = difference between the cloud base temperature and snow surface temperature, °F

N = estimated cloud cover expressed, decimal fraction

k = basin convection-condensation melt factor expressing average exposure to wind

g. Open-area equation. This equation is based upon theoretical principles, with coefficients determined on the basis of observations at a lysimeter at the Central Sierra Snow Laboratory. Shortwave radiation, usually always the most important melt factor in this setting, is based upon the measured or assumed incident solar radiation (taking into account cloud-cover estimates), together with snow surface albedo and the basin shortwave radiation melt coefficient k'. Long-wave radiation is calculated on the basis of the air temperature relationship (Equation 5-8) for cloud-free periods. For periods with cloud cover, Equation 5-9 is applied, using the difference between the temperature of the cloud base and the snow surface temperature. The cloud base temperature can be estimated from upper air temperatures or from lapse rates from a surface station. Convection and condensation, usually of relatively minor importance in this setting, are computed using Equation 5-15.

h. Partly forested area equation. This equation and those following reflect a different method of derivation from the procedures used for Equation 5-24 and for the rain-on-snow equations. Instead of relying entirely on a theoretical factors, multiple regression techniques employing field-laboratory data were used to establish some of the coefficients for a given forest cover. Thus, in the treatment of shortwave radiation and long-wave loss, the long-wave loss is computed indirectly by incorporating it into the statistically

derived shortwave radiation coefficient. This results in a coefficient of 0.0040 compared with the theoretical value of 0.00508. The shortwave radiation is computed only for nonforested areas, using the effective forest canopy factor *F*.

i. Forested-area equation. Equation 5-26 reflects the assumption that shortwave radiation is unimportant because of the forest cover. The basin-mean wind, however, is assumed to be significant enough to effect convection-condensation melt and is computed as in Equation 5-25. Long-wave radiation from the forest canopy is computed as a function of air temperature as in Equation 5-25.

j. Heavily forested-area equation. This equation is obtained from correlation analysis of data for the Willamette Basin Snow Laboratory, a heavily forested field site. (See Table 5-3 for specific references.) The convection melt term is $0.011(T_a - 32)$; the long-wave radiation term is $0.029(T_a - 32)$; and the condensation melt term is $0.034(T_d - 32)$. Combining these terms yields Equation 5-27.

5-5. Summary of Generalized Energy Budget Equations

Tables 5-2 and 5-3 summarize the energy budget equations for rain-on-snow and rain-free situations.

5-6. Sensitivity of Variables and Coefficients in Generalized Equations

a. Overview. This section discusses the relative magnitude of the snowmelt components described in Paragraph 5-3 and contained within the generalized equations presented in Paragraph 5-4, including further analysis of the sensitivity of the variables and factors inherent in the equations. The discussion is intended to assist the practitioner in applying either the energy budget equations or temperature index procedures (Chapter 6) for snowmelt analysis and simulation. The paragraph addresses questions such as which factors are most important in given meteorological and geographical settings and where emphasis should be placed in obtaining data and performing the analysis.

b. Relative magnitude of melt components. The energy budget equations provide a convenient means for illustrating the relative magnitude of the snowmelt components for specified conditions. In Table 5-4, seven assumed settings are postulated, together with the assumed meteorological conditions. Three are for rain-on-snow and three are for rain-free conditions. The melt quantities are computed using the appropriate generalized equation. For the rain-free condition, the melt quantities illustrate the importance of shortwave radiation as a melt-producing source and also show how cloud cover and albedo changes can significantly affect this melt component. For the rain-on-snow condition, the dominance of condensation melt can be seen, along with the importance that wind velocity plays in this component. Rain melt, by contrast, is relatively small, even for the condition having relatively heavy rainfall.

(1) Cases 1 and 3 illustrate the effects of cloud cover in a rain-free situation. Two factors are at work: first, shortwave radiation is reduced because of cloud cover, and second, net long-wave radiation is increased as outgoing radiation is decreased and back-radiation from clouds is increased. These two melt components therefore tend to offset themselves. This suggests that cloud cover is a somewhat insensitive variable in the overall equation once the maximum possible insolation rate is established for the time of year and latitude.

(2) Further illustration of the relative contribution from the energy budget components is shown in Figures 5-6 through 5-8. For these relationships, daily snowmelt has been computed from the appropriate generalized equation and plotted against air temperature as the main independent variable (x), with a second variable as a parameter (z). They illustrate the variability and importance of the second variable compared with the most frequently used index variable, air temperature. These plots will be referred to again in the discussion of the temperature index method (Chapter 6). Figure 5-6 shows that, during rain-on-snow, precipitation magnitude does not introduce a significant additional variance in melt over that supplied by air temperature. In Figure 5-7, wind velocity—affecting convection and condensation—is an important variable in computing snowmelt, having almost the amount of variance as temperature. Figure 5-8 shows the effect of wind velocity in a partly forested rain-free setting. Note the lower magnitudes of melt in comparison with Figure 5-7, because condensation melt is

Table 5-4
Magnitude of Melt for Identifiable Meteorological Settings

a. Assumed Conditions

Case	Description	Assumed Meteorological Conditions T_a	T_d	I_i	P_r	v
1.	Clear, hot, summer day. No forest cover. Albedo = 40%	70	45	700	0	3
2.	Same as Case 1, 40% forest cover	70	45	700	0	3
3.	Same as Case 1, 50% cloud cover	65	50	500	0	3
4.	Same as Case 1, fresh snow. Albedo = 70%	70	45	700	0	3
--	--------------------	--	--	--	--	--
5.	Heavy wind and rain, warm. No forest cover	50	50	0	3	15
6.	Same as Case 5, but light rain, windy	50	50	0	0.5	15
7.	Same as Case 6, but light wind	50	50	0	0.5	3

B. Daily Melt Quantities

Case	Snowmelt components, in. M_{sw}	M_l	M_{ce}	M_r	M_g	Total Melt in.	Rain + Melt in.
1.	2.13	-0.03	0.47	0	0	2.57	2.57
2.	1.01	0.44	0.28	0	0	1.73	1.73
3.	1.52	0.34	0.54	0	0	2.40	2.40
4.	1.07	-0.03	0.47	0	0	1.51	1.51
--	---	---	---	---	---	---	---
5.	0.05	0.52	2.27	0.38	0.02	3.24	6.24
6.	0.05	0.52	2.27	0.06	0.02	2.92	3.42
7.	0.05	0.52	2.27	0.06	0.02	1.11	1.61

Note: T_a = Air temperature, °F; T_d = Dew-point temperature, °F; I_i = Solar insulation, langleys; P_r = Daily rainfall, in.; v = Mean wind velocity, mph; M_{sw} = Shortwave radiation melt; M_l = Long-wave radiation melt; M_{ce} = Convection/condensation melt; M_r = Rain melt; M_g = Ground heat melt.

not as significant a factor in these dry conditions, and convection is a relatively small component.

c. Sensitivity of coefficients. In the previous discussions introducing the energy budget equations, the basis for the various factors and coefficients in the equations have been explained. Some are based upon theoretical principles; others are strictly empirical and perhaps vary a great deal. The degree of influence on the final outcome of the equation largely depends, of course, on the importance of the variable in the equation with which a coefficient or constant is associated. A summary of the most important and critical factors to be concerned about is listed below and are also noted in Tables 5-2 and 5-3.

(1) In the equations that take into account wind velocity in computing convection-condensation melt, the factors associated with the term become quite sensitive in influencing the computed melt. In a partly forested rain-on-snow situation, for instance, the convection-condensation term carries the most weight in determining melt, over 50 percent when wind velocity is relatively great. This places considerable importance on the wind exposure constant k, which can have a wide range of variation. As previously noted, that part of the coefficient 0.0084 pertaining to convection melt is also subject to wide variation as an experimental coefficient. These factors, therefore, can be considered to be sensitive and should be treated with care if they are subjectively determined. This concern can be reduced when using a model that can be calibrated and verified with historical data.

(2) In the rain-free equations, the dew-point variable is often not directly available and might be estimated on the basis of assumptions of relative humidity magnitude

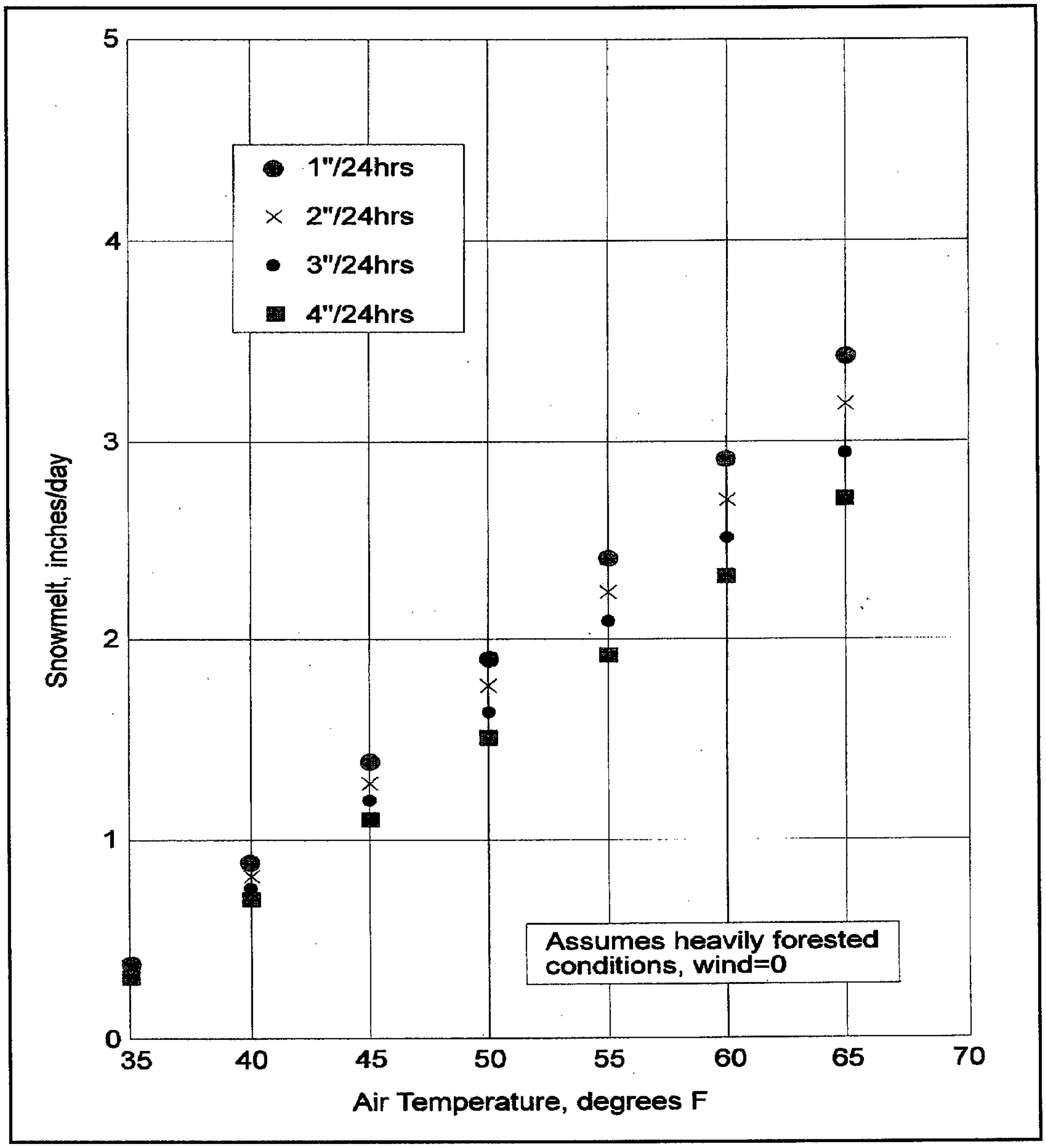

Figure 5-6. Melt versus temperature and precipitation, rain-on-snow area

and air temperature (Appendix D). Since condensation melt, which this variable indexes, can be one of the more influential components in computing melt, the dew point must be carefully estimated.

(3) In the equations that use solar radiation as a variable, the solar radiation term often becomes the most significant term in the melt equation. Thus, the factors k and F become relatively important, as does albedo. The factor k (shortwave radiation melt factor) is defined as being relatively insensitive, varying between 0.9 and 1.1. The forest-canopy cover factor F is a measurable factor that is therefore limited in its variability. The snow surface albedo, which must be calculated or estimated, can be quite variable in real-time during periods of snow accumulation, but should follow a relatively predictable decay function once snow ablation is underway. This factor is quite significant in affecting solar radiation melt magnitude as

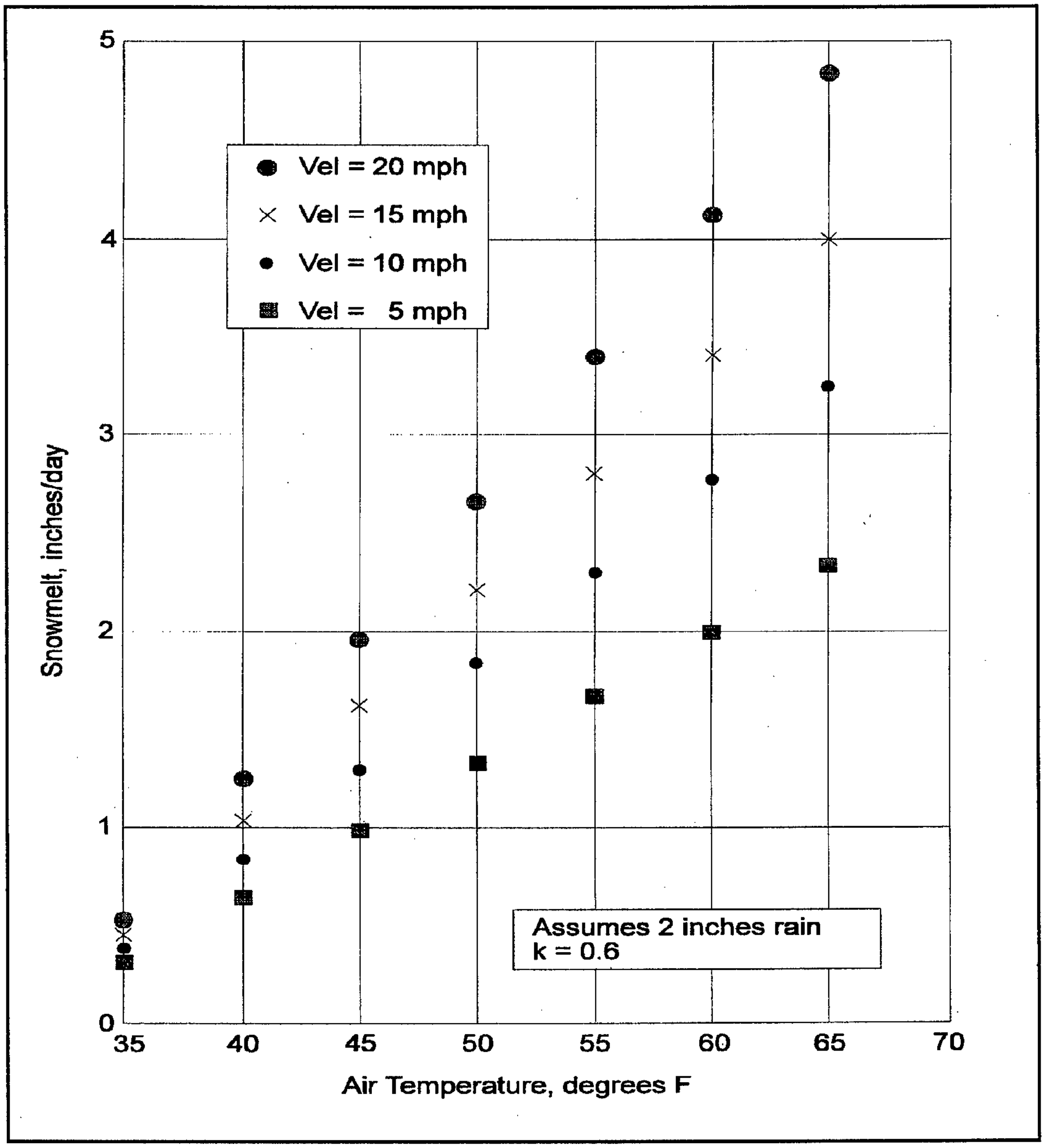

Figure 5-7. Melt versus wind and temperature, rain-on-snow area

demonstrated in Table 5-4. The coefficient 0.0040 in the partly forested equation has been determined by statistical means and, as discussed in *Snow Hydrology*, appears to have shown relatively good consistency when computed from different laboratory data.

(4) In general, care must be taken in choosing one equation over another on the basis of forest cover. A borderline forest-cover percentage could lead to quite different melt quantities, depending upon which equation was applied.

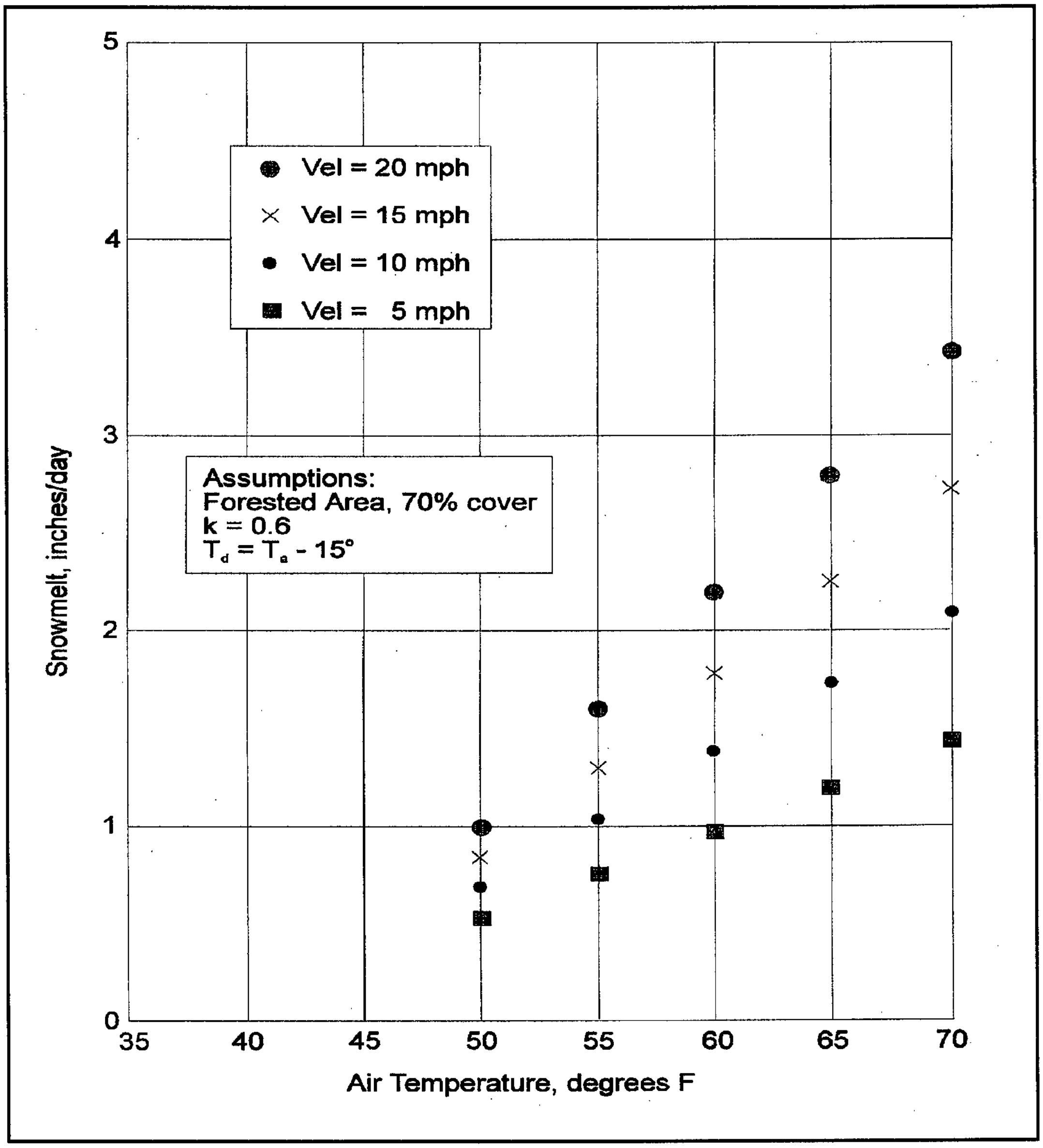

Figure 5-8. Melt versus wind and temperature, rain-free, partly forested area

Chapter 6
Snowmelt—Temperature Index Solutions

6-1. Overview—Basic Assumptions

This chapter covers the second basic method for computing snowmelt, that of using air temperature as an index to melt. This method recognizes the basic problem in applied snow hydrology, particularly in river forecasting, that many of the energy budget variables are not conveniently available for use. It also fully employs the concept of an "index," where a known variable is used to explain a phenomenon in a statistical rather than in a physical sense. As noted in Chapter 5, the snow investigations studies used the index concept for some of the energy budget equations by employing multiple regression and by simply accepting the fact that the physics involved were not explicitly explained in the parameters so derived. These statistical studies went further to explore the possibility of removing many of the "difficult" variables from the equations to make them as practical as possible. Since air temperature was already a predominant variable used in the energy budget equations, it is logically connected with many of the energy exchanges involved in snowmelt. And since it is commonly available to hydrologists in historical and real-time databases, the studies concluded that air temperature is a useful index to snowmelt, particular in forest-covered basins. Since that time, the temperature index method has been used extensively and almost exclusively in snowmelt modeling and river forecasting. This chapter will present the basic temperature index equation and technique, concentrating on the melt-rate coefficient, which is the key to using this approach successfully. The method will be compared with results achieved with the energy budget equations to illustrate the problems in applying the equation using nominal melt-rate factor values to situations beyond the boundaries of ideal application. Since this is a solution that needs to be applied with considerable judgment, summary guidance on the approach will conclude the chapter.

6-2. Basic Equation

The basic equation for the temperature index solution is

$$M_s = C_m(T_a - T_b) \tag{6-1}$$

where

M_s = snowmelt, in. per period

C_m = melt-rate coefficient that is often variable, in./(degree/period)

T_a = air temperature, °F

T_b = base temperature, °F

a. In the above equation the melt-rate factor C_m typically varies between 1.8 and 3.7 mm/°C (0.04 and 0.08 in./°F) as discussed in detail below. The temperature variable used would depend upon the method of application and the size of the river basin involved. For large snowmelt basins simulated with a daily time increment, it is typical to use daily maximum and minimum air temperatures as the index variables for this equation, weighting each as desired based upon model calibration. Sometimes the maximum daily temperature only is used as the index because it is an indicator of cloud cover in the basin. If the computation interval needs to be shorter than 1 day, then representative average temperatures for the computation period would be used.

b. The base temperature is typically a value near 0 °C (32 °F), particularly for shorter computation periods using representative period temperatures as input. If, however, maximum daily temperatures were used as the index, the base temperature would be higher, perhaps as high as 4.44 °C (40 °F).

c. Investigators have over the years offered variations to Equation 6-1, primarily in the manner of specifying the melt-rate factor. Gray and Prowse (1992) contains a good summary of these alternative expressions.

6-3. Melt-Rate Coefficient—Sensitivity, Range of Magnitude

a. General. Proper use of the melt-rate coefficient (sometimes called the degree-day factor) is an important key to successfully applying the temperature index equation. Review of the discussion in Chapters 2 and 5 of the physical principles involved in snowmelt shows intuitively that temperature is not

directly related at all to shortwave or condensation energy sources, and it only partially explains the other components of total energy flux. In the derivation of the generalized energy budget equations, however, it was demonstrated that, for forested areas, shortwave radiation and wind effects become less important, thereby allowing temperature to become a more definitive index of snowmelt. In general, then, it can be said that temperature is a reasonably good index of energy flux in heavily forested areas, while it is less so in open areas where shortwave radiation or wind velocity plays a more important role in the melt process. It follows that for those situations where the factor is not a good index, the melt-rate factor must be treated less as a constant and more as a variable to make the application work with reasonable success. This is accomplished by having the melt-rate factor vary according to independent relationships in a simulation model or by simply applying careful judgment in choosing the appropriate value, say in a river-forecasting situation.

b. Range of variation. The range of the melt-rate factor is typically 1.8 to 3.7 mm/°C (0.04 to 0.08 in./°F) for rain-free conditions. Higher values can be expected in extreme cases, as will be demonstrated. These factors would be lower if the temperature index used is the maximum daily temperature. The possible range of the melt-rate factor can be illustrated by the hypothetical cases presented in Table 5-4. By use of the daily melt quantity calculated by the energy budget equations and the temperatures assumed, the melt-rate coefficients calculated through Equation 6-1 would be as shown on Table 6-1. This table demonstrates that one case where the nominal values of melt-rate coefficient would underestimate snowmelt is when heavy winds in a rain-on-snow situation with a saturated air mass cause condensation melt to be high. Additionally, Figures 5-6, 5-7, and 5-8 and overlying lines representing the temperature index equation with varying melt-rate factors illustrate the melt-rate factor range. These are shown in Figures 6-1, 6-2, and 6-3. In general, these suggest that a melt-rate factor for rain-on-snow should be on the high side of the nominal range, and, for situations where wind is import, even higher values should be used.

Table 6-1
Relative Magnitude of Melt-Rate Factors (Refer to Table 5-4)

Case	T_a, °F	T_b, °F	Melt, in.	C_m, in./°F	Comment
1	70	32	2.57	0.068	Clear, low albedo
2	70	32	2.40	0.073	Case 1, 40% forest
3	65	32	1.51	0.040	Case 1, cloud cover
4	70	32	1.73	0.046	Case 1, fresh snow
---	---	---	---	---	-------------
5	50	32	3.24	0.180	Heavy rain, windy
6	50	32	2.92	0.163	Light rain, windy
7	50	32	1.11	0.062	Light rain, light wind

c. Determination and application. In modeling for engineering and forecasting practice, the melt-rate factors are verified through the process of calibrating a hydrologic model. The energy budget equations can be a useful guide in establishing initial estimates for the model. Once established for known historical conditions, the factor can be modified by judgment. Again, the energy budget principles should be applied in assisting in this process. Additional discussion of the magnitude of the temperature index melt-rate factor can be found in USACE (1956), Anderson (1973), and Male and Gray (1981).

(1) Real-time flood forecasting in some rain-on-snow situations may present challenges in using the temperature index method since, as shown above, the melt-rate coefficient can vary widely in magnitude because of wind effects. In major storms, the variation could be abrupt and have quite a significant effect on snowmelt rates. If this is a potential problem, real-time wind data should be used. If not factored directly into a simulation model, the wind data could be used with a relationship, such as that shown in Figure 6-2, as guidance to a forecaster who is making on-the-spot-judgment calls in setting up a forecast model. The relationship should be verified with known historical storm data, if possible.

(2) In clear-weather and partly forested snowmelt situations, the melt-rate factor varies seasonally, typically increasing as the snowmelt season progresses owing to factors such as the decrease in albedo and

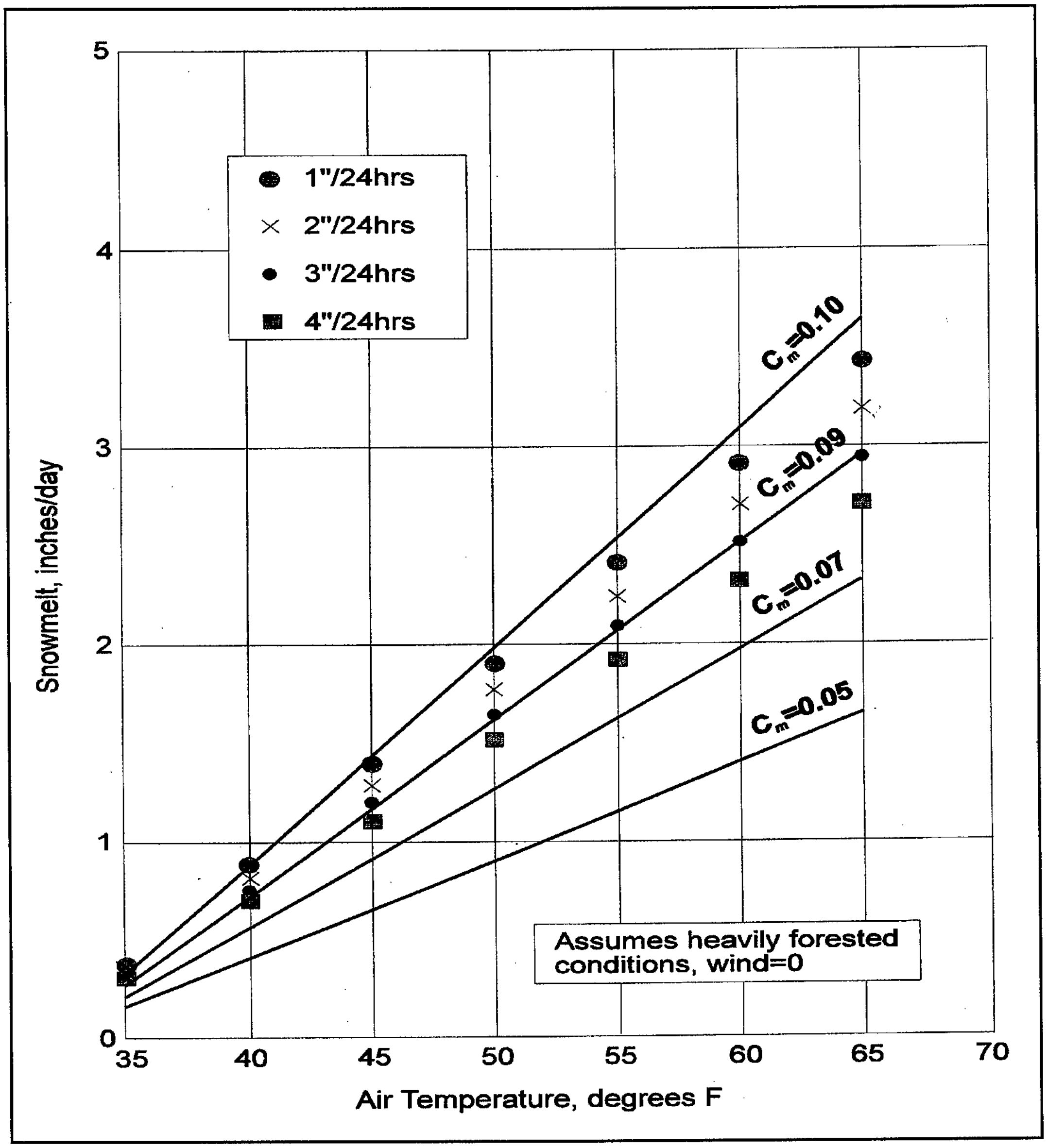

Figure 6-1. Melt rates in a heavily forested area, rain-on-snow

increased daily insolation. Because of this, simulation models usually calculate C_m as a variable. This can be done by making C_m a function of accumulated runoff or accumulated degree-days of air temperature. Such a relationship would need to be verified by simulation of historical records.

(3) For hydrological engineering analyses, the melt-rate factor must be used with considerable caution, if at all. In a well-calibrated and verified model, the factor is not a concern unless the design application extrapolates beyond the range of historical calibration. In such cases, reference to the energy budget equations may help in judging the magnitude of the melt-rate factor. For derivations of extreme floods, such as a PMF, the temperature index approach is of little or no value in computing snowmelt-rates, since there is no way to quantify a maximum snowmelt. In such applications, the energy budget method should be used.

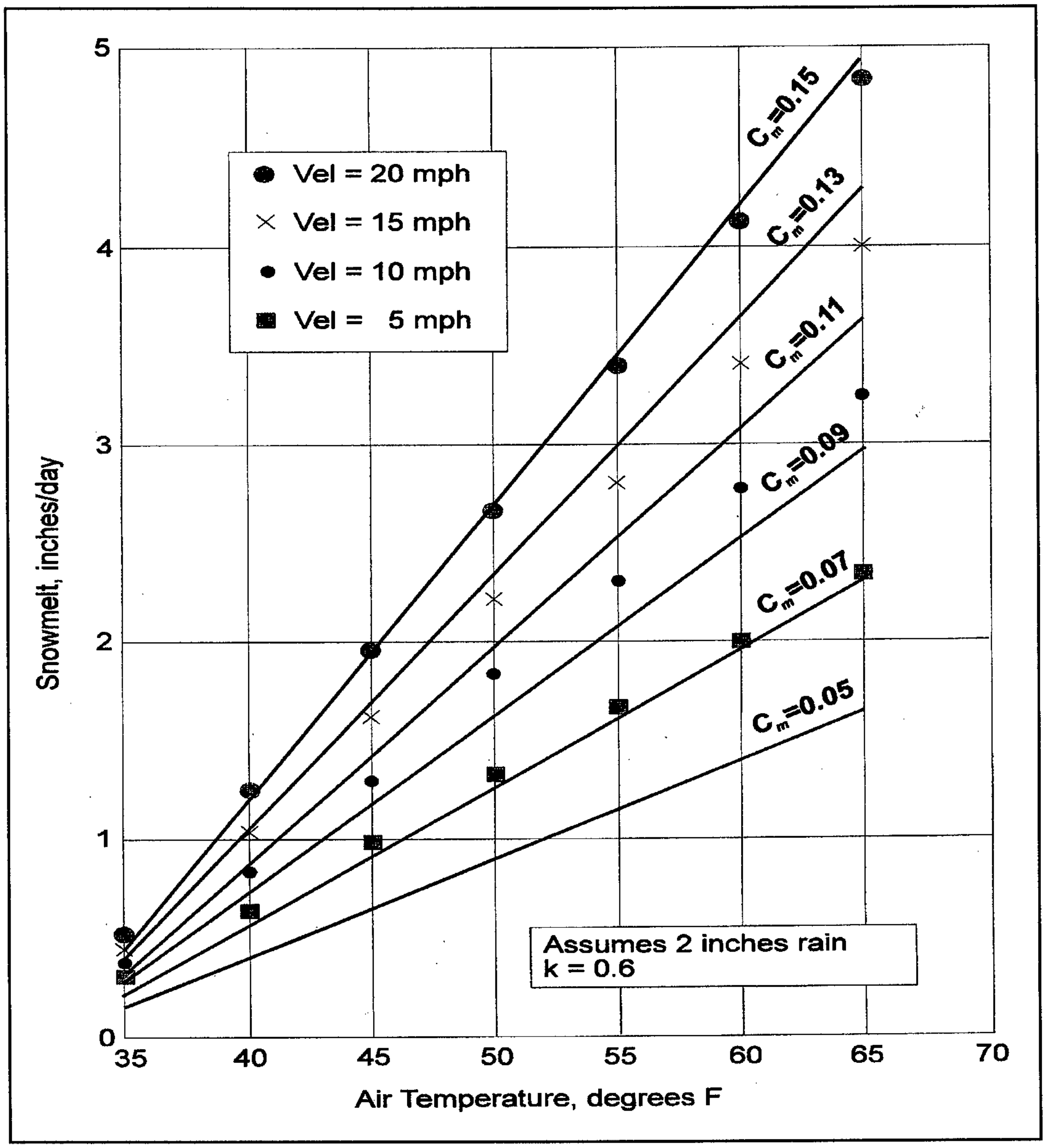

Figure 6-2. Melt rates in a partly forested area with wind effects, rain-on-snow

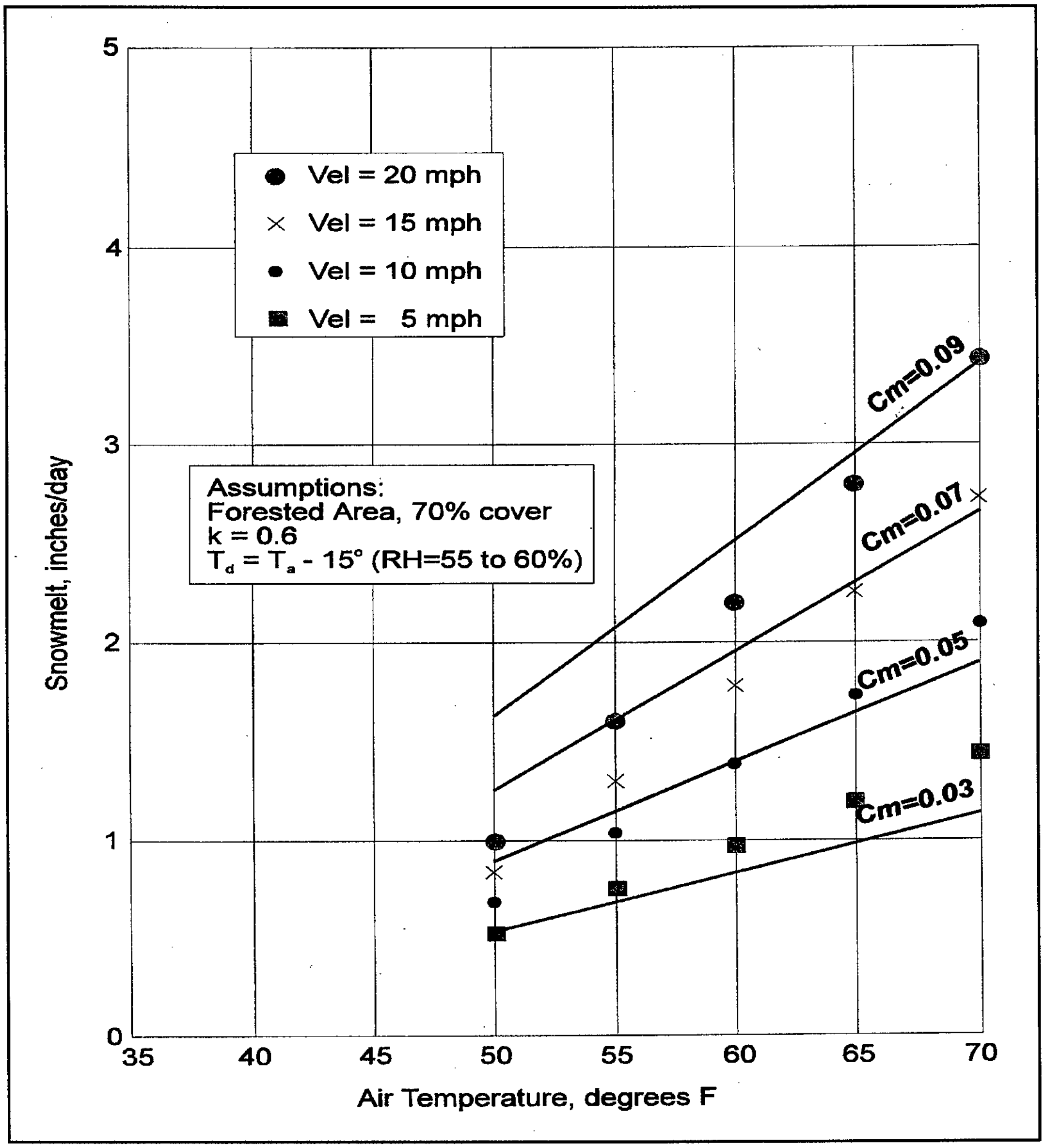

Figure 6-3. Melt rates in a forested area, rain-free conditions

Chapter 7
Effect of Snow Condition on Runoff

7-1. Overview

This chapter is primarily about the state of the snowpack during the winter accumulation period and into the early spring and the effect that the snowpack has on delaying runoff from rainwater and melt during this time. The techniques that will be described are generally applicable to rain-on-snow conditions frequently experienced in basins of the eastern and western United States that are subject to winter maritime rainstorms. Hydrological analysis or forecasting under these conditions requires a particular awareness of the ability of the snowpack to store water, thus delaying runoff to some extent. The magnitude of this effect will be discussed, and methods for determining and simulating the storage effect of the snowpack for practical forecasting and design will be presented.

a. Chapter 2 has described the changes in the character of the snowpack as it is transformed from a fresh, low-density, crystalline state to a dense, coarse-grained condition that is isothermal at 0 °C and ready for melt. In a rain-on-snow environment, these conditions are particularly dynamic, continually changing as the basin is subjected throughout the winter to a succession of storms—bringing precipitation either in the form of rain or snow—interspersed by dry periods that are often below freezing at higher elevations. This changing environment must be considered in analysis and modeling, and the changing character of the snowpack as it affects runoff must be a part of continuous simulation models. The following phenomena must be considered in one way or another.

(1) As rainwater or melt enters a subfreezing snowpack, it must first give up energy to raising the temperature of the snow before it can be available for runoff.

(2) In addition to the rainwater and melt that is frozen in the snowpack, an additional amount is lost in satisfying a liquid-water capacity that is inherent in fresh snow.

(3) In the process of traveling through the snowpack, the rainwater and melt may follow a circuitous route as it encounters ice lenses and "cold" pockets within the snowpack, thus delaying the delivery of water to the ground surface.

b. A wide spectrum of alternatives is currently employed in practice in dealing with the above factors, ranging from detailed physical modeling of the snowpack's internal characteristics to simply considering these effects to be small enough that they can be ignored. Fortunately, for some engineering applications the snowpack's condition can be ignored. In design flood derivations, for instance, the snowpack can be assumed to be fully ripe before the flood begins; in many forecasting settings, the uncertainty of many other factors often outweighs the relatively small magnitude of snow-condition effects. In this chapter an overview will be given on the approaches to modeling the condition of the snowpack, and a discussion on the relative magnitude of the phenomena will be presented.

7-2. Cold Content

For practical applications, the concept of cold content is used in quantifying the effect of the snowpack temperature on rain and melt. Cold content defines the amount of energy needed to raise a "cold" snowpack to 0 °C, expressed in terms of the amount of water needed to be produced at the surface to release energy by freezing. This can be calculated by:

$$W_c = \frac{0.5\,\rho_s d T_s'}{80} \tag{7-1}$$

where

W_c = cold content, in.

0.5 = specific heat of ice, cal/g °C

ρ_s = average density of the snowpack, g/cc

d = depth of pack, in.

T'_s = average temperature deficit of snowpack below 0 °C, °C

80 = latent heat of fusion of water, cal/cc

For practical applications, the average snowpack temperature can be estimated on the basis of the air temperature for 1 to 3 days before the forecast time. Typically, the temperature will be close to that of the air at the snowpack surface, but will approach 0 °C at the ground. For deep snowpacks, a further assumption can be made that only the top 61 cm (24 in.) or so of snow is subject to the influence of air temperature and that the deeper pack is only 1 to 2 degrees or so below freezing. The density of this layer of snow can also be assumed to be greater than the top layer. Examples of computation are presented later in this chapter.

7-3. Liquid-Water-Holding Capacity

As summarized above, the liquid-water-holding capacity of the snow is a second factor that can be considered an "initial loss" in practical applications in snow hydrology and forecasting. Unfortunately, there is very little experimental evidence leading to the quantification of this. It varies, depending upon the depth and density of the snow, the mass of ice layers, and the channelization and honeycombing of the snowpack. At 0 °C this factor is approximately 2 to 5 percent of the SWE (USACE 1956). For most practical applications, a fixed percentage of the SWE is used as an initial loss, in addition to the cold-content loss. Note that this magnitude of loss assumes the free drainage of the water. Therefore, in flat areas the snowpack may hold liquid water far in excess of the amount that is found in mountainous areas.

7-4. Movement of Water Through the Snowpack

The final effect of a snowpack on rainwater and melt is the time delay as liquid water moves downward to the ground surface. This process has been explored in laboratory experiments, as discussed in Chapter 2, and theoretical equations have been developed to explained the phenomenon. Anderson (1973) has developed empirical relationships using time lag and attenuation to represent drainage through a snowpack. However, in practical applications, this seldom is considered a significant enough delay to warrant a detailed evaluation. The snow investigations studies noted that the net storage effect on water draining through a moderately deep snowpack resulted in a time delay on the order of 3 to 4 hr.

7-5. Simulating Change in Snow Condition

Simulation of the above phenomena involves the following considerations:

- Calculating the gain or loss of heat from the air, liquid water, and ground sources.
- Maintaining a continuous accounting of negative heat storage in the snowpack, including diurnal.
- Maintaining an estimate of the snowpack thermal conductivity, a function of snowpack density.
- Accounting for the variation in snowpack character in the vertical dimension.
- Keeping an accounting of the liquid water currently in the snowpack, in both the retained and gravitational phases.
- Estimating the attenuation and time lag of gravitational water movement through the snowpack.

Many of these processes are obviously complex and therefore are computed explicitly in only detailed physical models. In a physically based simulation, an internal mass balance is continuously computed as a part of the basic energy balance of the snowpack (Equation 2-1). Snowpack settling and density may be continuously estimated, with the snowpack definition being accounted for in more than one vertical layer.

a. An empirical approach that is currently used widely is that of Anderson (1973). Here, an accounting is maintained of the relative temperature of the snowpack below freezing as a function of time. In effect, the snowpack is simulated as an energy reservoir; once the reservoir is full (snowpack isothermal and at 0 °C) meltwater moves to the ground. This can be done through an index relation such as:

$$T_s(2) = T_s(1) + F_p(T_a(2) - T_s(1)) \qquad (7\text{-}2)$$

where

T_s = index of the snowpack surface temperature at times (1) and (2)

F_p = factor, varying from 0 to 1, representing the relative penetration of the air temperature into the snowpack

T_a = air temperature

b. If F_p is close to 1.0, the snow temperature will remain close to that of the air; thus, values close to 1.0 would be appropriate for shallow snowpacks. For a deep snowpack, a low value of F_p will result in a slow cooling or warming of the snow. The variable T_s is limited to a maximum of 0 °C (32 °F) in the simulation process.

c. Once a snow-surface temperature index is established for a computation period, the cold content can be calculated through an equation such as:

$$W_c(2) = W_c(1) + C_r\,(T_a(2) - T_s(1)) \qquad (7\text{-}3)$$

where

W_c = cold content, in. (mm)

T_a = air temperature

T_s = index of the snowpack surface temperature

C_r = conversion factor, in. (mm)/degree-day

The value of C_r can be made a variable in simulation models by relating it to calendar periods or to a cumulative temperature index function. Figure 4-1 is an example of a computer printout made during a winter-snow accumulation period. In this model, snow conditioning is simulated using the above technique. Note that for several periods following a cold period, snowmelt is limited by satisfying cold content and liquid-water deficiency requirements.

7-6. Impact on Runoff

The magnitude of cold content can be illustrated by calculating this factor for various assumed conditions using Equation 7-1. This is shown in Table 7-1. For the deep snowpack example, the calculation is subdivided into two layers, above and below a 61-cm (24-in.) depth. As can be seen, the cold content is a relatively small factor compared with the potential magnitude of rainfall and snowmelt. It varies from 3 percent or more of the SWE for "cold" snow, to 1 percent or less of the SWE for deeper snow that is closer to 0 °C.

a. A second illustration, based upon observations made at the Willamette Basin Snow Laboratory, is shown on Figure 7-1. Illustrated is the storage and transmission of water in the snowpack for an observed rain-on-snow situation. Here, a snowpack having a water equivalent of 67.8 cm (26.7 in.) receives input from a 2-day rainstorm. Since the snowpack was colder than 0 °C (at -6 °F), some rain and condensation was taken up in raising the temperature of the snow to 0 °C through the process of freezing; this amounted to 0.76 cm (0.3 in.) of water. An additional amount of liquid water, 1.27 cm (0.5 in.) was permanently retained in the snowpack, making the total quantity of stored water 2 cm (0.8 in.) Finally, the snowpack also temporarily stored some water input as it progressed through the pack. In this case, given the rate of input involved, there was a time delay of 12 hr between the beginning of rain and melt and the beginning of runoff. The hydrograph resulting from the input summarized in Figure 7-1 is shown in Figure 7-2. The loss of 2 cm (0.8 in.) is displayed.

b. For practical applications in a rain-on-snow situation, snow-condition effects can be thought of as an "initial loss" that is subtracted from input, much in the same way as initial losses in dry-soil conditions are simulated in rain-runoff situations. For the engineer, the problem is to be able recognize this potential and to be able to incorporate this time lag in the snow hydrology analysis, where appropriate. Practically speaking, this may not a major factor in design analysis since the snow can usually be assumed to be fully primed prior to the beginning of significant runoff producing melt. In certain forecasting situations, however, the effect of snow conditioning can be noticeable, and it is definitely a factor that needs to be considered in continuous simulation models that operate through periods antecedent to active snowmelt periods.

Table 7-1
Variation in Cold Content

	Assumed Factors			Calculated Factors		
Descriptive Condition	***d*, in.**	**ρ**	**T'_s, °C**	**SWE, in.**	***Wc*, in.**	***Wc*/SWE, %**
Shallow, relatively fresh snowpack. Several days of 8 °C temperatures prior to application	16	0.20	6.0	3.2	0.12	3.8
Same, but warm snowpack	16	0.20	1.0	3.2	0.02	0.6
Deep snowpack, top 61-cm (24-in.) layer cold	24 36 60	0.20 0.30	5.0 1.0	4.8 10.8 15.6	0.15 0.07 0.22	1.4
Deep, ripe snowpack. Warmer antecedent conditions	24 56 80	0.35 0.45	1.0 0.5	8.4 25.2 33.6	0.05 0.08 0.13	0.4

Note: d = snow depth, in.; ρ_s = snow density, g/cc; T'_s = average temperature of snow layer, °C below freezing; SWE = beginning snow-water equivalent, in.; W_c = cold content from Equation 7-1, in.

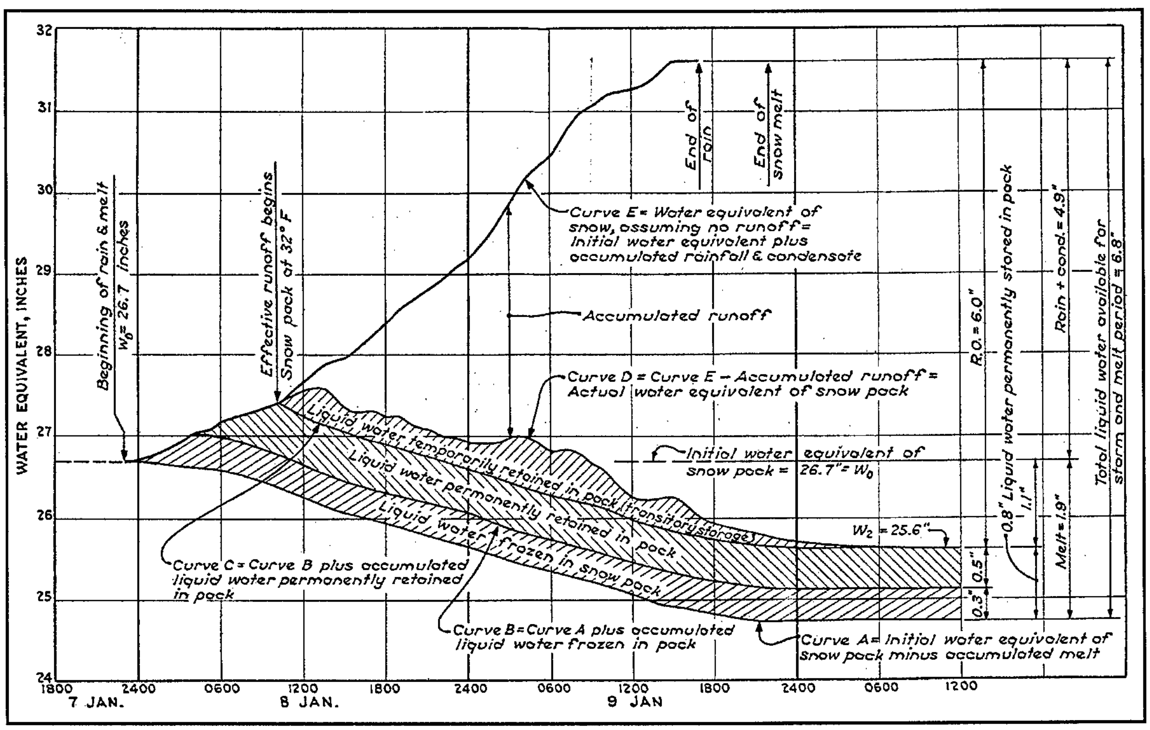

Figure 7-1. Snowpack water balance during rain on snow (Figure 4, Plate 8-9, *Snow Hydrology*)

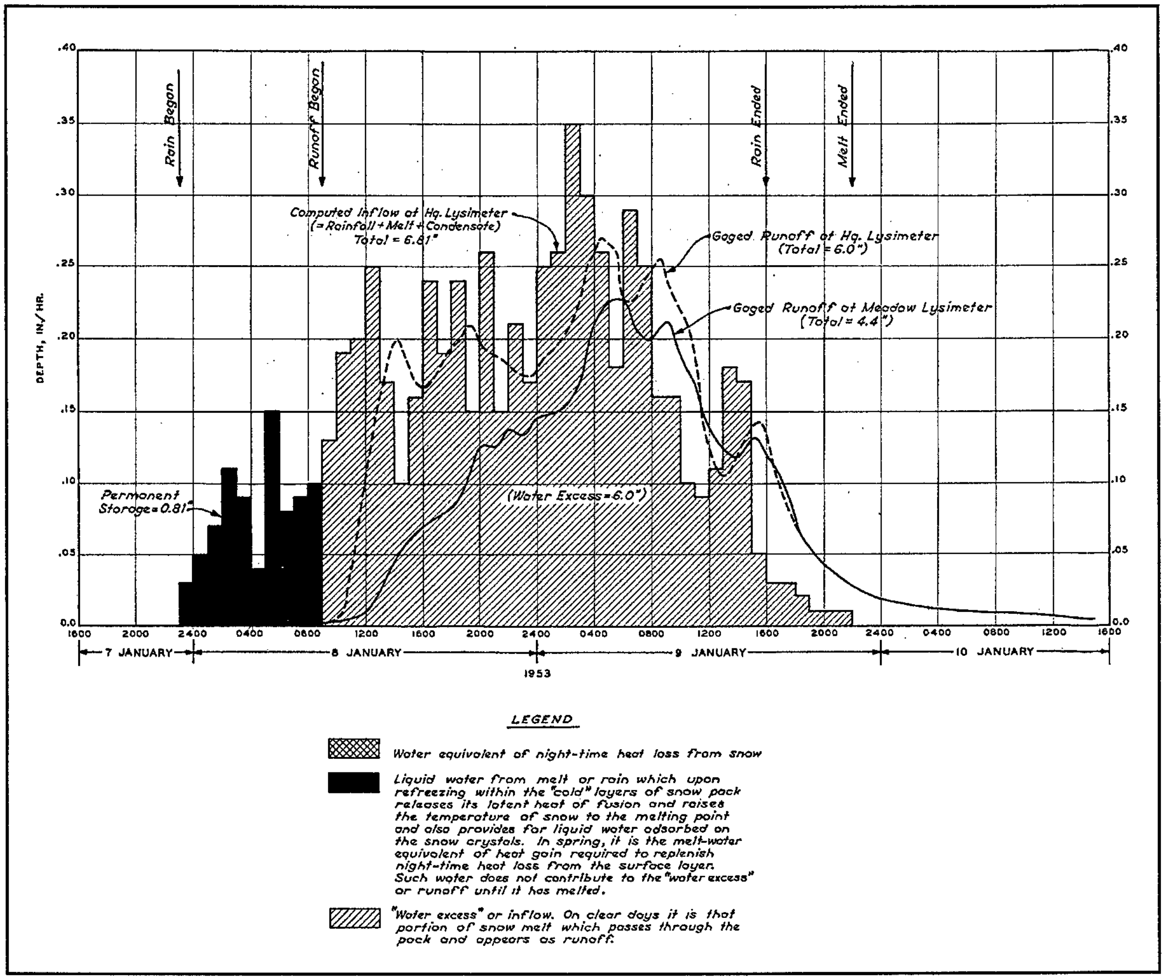

Figure 7-2. Hydrograph resulting from Figure 7-1 rain-on-snow event (Plate 8-8, *Snow Hydrology*)

Chapter 8
Snowmelt—Accounting for Changes in Snow and Snow Cover

8-1. Overview

This chapter describes the requirements needed and techniques used to track the state of the snowpack in a basin once the accumulation of snow has ended and ablation has begun. It follows logically in sequence after Chapter 7, which has covered the internal changes in the snowpack, primarily in winter in rain-on-snow situations or early spring and how they affect snowmelt. This discussion is oriented primarily to spring-summer snowmelt in the large interior basins of the western United States, where snowmelt is a 2- to 3-month-long process. The following is a summary of the changes that take place in the snowpack and its watershed during snowmelt for spring-summer:

- The snowpack, now internally isothermal and at 0 °C, yields meltwater to the soil surface as heat energy is applied at its surface and ground.

- The snow surface albedo continues to decline as surface snow crystals become rounded. This allows greater amounts of shortwave radiation to be absorbed as heat energy.

- As snow melts, first at lower elevations, the snowline begins to climb to higher elevations. This shifts the melting level in the basin to higher and higher elevations as the season progresses.

- As the snowpack recedes, the snow-covered area of the basin decreases, while the snow-free area increases. The soil moisture in the snow-free area decreases, thereby leaving the basin with two distinctly different runoff characteristics.

- Any precipitation falling during the melt season will encounter a variety of potential situations: it will fall as fresh snow at higher elevations, as rain-on-snow at lower elevations, and as rain on bare ground (with reduced soil moisture) at low elevations.

- As the melt season progresses to its later stages, the active melt zone may shift from a forest-covered area to one that is free from forest cover, above the timberline. This results in new energy sources dominating the snowmelt process.

a. The problem for forecasting and analysis is not only to account for the above phenomena if they are important in a particular application but also to accurately as possible assess the residual SWE or volume of runoff anticipated. An initial volume of SWE is determined at the beginning of the snowmelt period, as discussed in Chapter 4. As the melt season progresses, calculated melt is subtracted from the initial values to yield a residual, and any additional precipitation is added. Any error in the initial estimate is carried into the residual; as the residual decreases, the error becomes more and more significant. This calls for the ability to update the residual snow-runoff estimate carried in the model by checking with observations in the basin.

b. In rain-on-snow situations, the meteorological conditions are such that most of the phenomena described above have little relevance. Here, the freezing level is continually shifting with the passage of storms, and the watershed's soil moisture may be saturated by rainfall, whether on snow or not. A snow-covered area may change in a matter of hours, rather than weeks, during a significant storm. The magnitude of the snowpack volume may be relatively small compared with the rainfall runoff involved. Solar radiation is often of little consequence. Despite the differences, however, modeling in this environment still requires the accounting of the snowpack during melt. Often, short-cuts and subjective methods are employed in operational applications.

8-2. Simplified Methods, Lumped Models

a. Simple estimates. The simplest approach in dealing with changes in the snowpack during melt is to assume that the changes are insignificant. This may be a reasonable assumption for rain on snow. If, for instance, the rainstorm is relatively short and the snowpack large, there may literally not be any change in the snowpack's areal extent during the storm. Chapter 10 discusses this further in conjunction with

design flood analysis. For river forecasting during rain on snow, manual updates, based upon real-time observations, will help determine the status of the snowpack. A further check in forecasting is to see how well the model is tracking observed streamflow.

b. Snow cover depletion curve. An approach that has been used in lumped models for spring-summer melt settings is to employ a snow cover depletion curve that describes the basin's snow-covered area as a function of accumulated snow runoff. Used in conjunction with an area-elevation curve, the snowline elevation for the basin can also be determined. An example of a generalized depletion curve as used in the SSARR model is shown on Figure 8-1. The "theoretical depletion curve" is derived using historical field and remote-sensing records together with runoff data. Studies have shown that this generalized relationship is relatively uniform for a basin. Observed conditions of snow cover and runoff, however, may yield a point that is not on the theoretical curve. In this case a proportionally adjusted curve is followed, as shown in Figure 8-1.

(1) While the snow cover depletion curve yields an accounting of the snow cover, this method still needs to independently estimate expected total basin SWE. The typical approach is to use multiple-regression procedures as described in Chapter 9 to determine an initial estimate of the total SWE (actually, expected total basin seasonal runoff). The accounting of currently remaining SWE during the melting of the snowpack is simply a process of subtraction. Adjustments in expected residual runoff and snow-covered area are periodically made during the snowmelt season using satellite data and fixed-wing reconnaissance flights and by verifying model performance by comparing observed and computed streamflow. This methodology is used in the Snowmelt Runoff Model (SRM) (Chapter 11).

(2) A consideration with this type of approach is how to compute runoff from the snow-free portion of the basin during spring or summer rain. One option that has been successful in the Columbia basin is to simply assume that summer rain falling over the snow-free area is negligible, since soil moisture is relatively

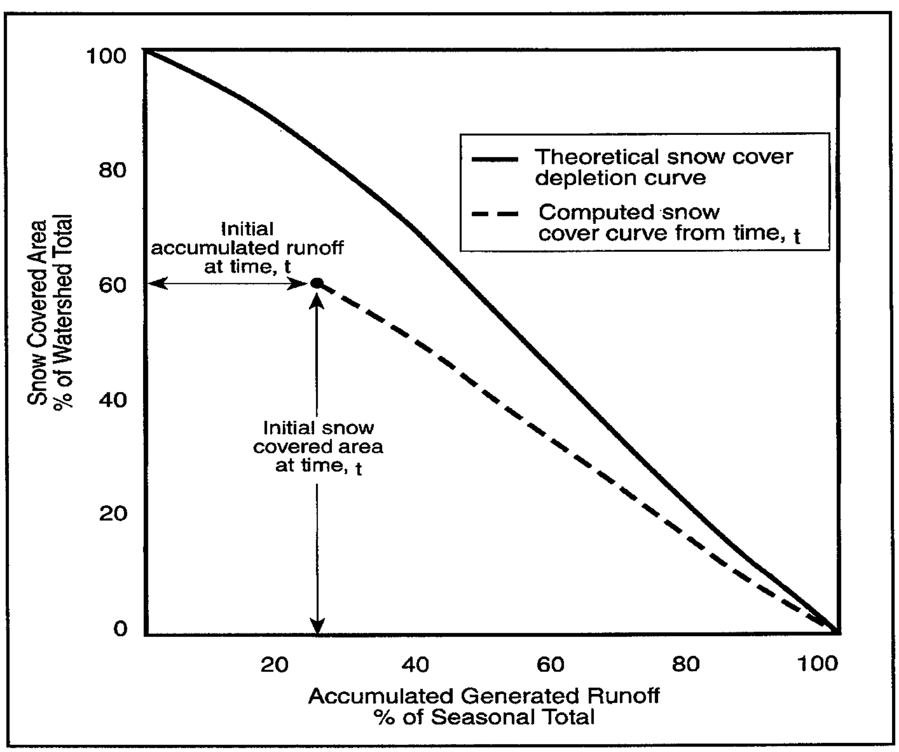

Figure 8-1. Example of snow cover depletion curve

low and rainfall quantities are not normally great. An alternative to this is to split the basin into a snow-covered and a snow-free zone. The snow-covered area is continuously defined with a snow cover depletion curve, and the snow-free component is computed as a complement to that area. With this technique, runoff from the snow-free portion is independently computed and added to the snow-covered runoff. Both of these options are available in the SSARR program under the lumped basin options.

(3) The snow cover depletion curve method is suitable for some design flood derivations in summer snowmelt settings since the depletion curve is based upon historical conditions, and initial SWE can be determined by independent analysis of historical records. This approach may not be valid, however, if the design condition includes a heavy spring rainstorm in addition to snowmelt.

8-3. Detailed Methods, Distributed Models

With all of the changes in the watershed and in a snowpack taking place during snowmelt, simplified approaches are limited in their ability to address many of these changes. A distributed model is required to begin accounting for changes in any detail.

a. Snow-band formulation. This method of defining a basin model, described in Chapter 4, can be employed with reasonable success to account for changes in the snowpack. The accounting of snow quantity, cover, and quality is done zone by zone. There is no reason why this cannot account for all the physical changes that occur during snowmelt. An important consideration, however, is whether each zone is assumed to be either 100-percent snow-covered or snow-free. If so, the basin may require a large number of zones to be adequately represented. Even with a large number of zones, the snowline can abruptly change as a zone transitions from being snow-covered to snow-free, causing unrealistic results in simulated flow. Because of this, a model may allow simulation of a gradual transition in snow cover within a zone. Figure 8-2 is a portion of summary printout from an SSARR model simulation, showing the changes in snow cover on eight bands of elevation. In this model, snow conditions are strictly homogeneous on each band, but a limiting function prevents abrupt transitions as a band becomes depleted of snow. An indicator flag shows when this has happened.

b. Grid-cell-based models. As with the snow-band approach, a horizontally defined grid system can also account for changes in the snowpack, provided the grid is fine enough. The same problems crop up as in the elevation band definition if homogeneous conditions are assumed and abrupt transitions occur.

8-4. Snow Observations During Snowmelt Forecasting

Regardless of the simulation technique used during snowmelt, an essential operational practice for runoff and streamflow forecasting is to make use of field observations to verify the model's state variables. This can range from simple subjective checks, based upon a limited amount of data, to complex systematic procedures. Three methods of employing field data are summarized below. This subject is discussed in more detail in EM 1110-2-9038.

a. Areal snowcover. This is a parameter that is fairly easily obtained, either from satellite imagery or by special aerial reconnaissance flights. Rango and Itten (1976) have effectively employed satellite observations in accounting for snow during snowmelt. The National Weather Service's Remote Sensing Center in Minneapolis has an ongoing program of providing processed snow-cover data to cooperating agencies during the spring-summer snowmelt period in the western States.

(1) An older approach still used by some USACE offices is flying fixed-wing aircraft into the basin at or near the snowline elevation and reporting the status of the snowline at fixed reference points. These data are converted to snow cover using an area-elevation curve. Snow-flight data are now used where satellite data are not yet satisfactory, or to simply to augment the remotely sensed data.

(2) Both the satellite observations and aerial reconnaissance can be obscured by a cloud cover. With satellite passes being at fixed intervals, it is possible to miss having snow cover information for an extended period. Partial cloud cover can be

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                                   SSARR SNOWBAND MODEL (METRIC) -
      COMPUTED FLOW, ILLECILLEWAET R., CANADA
           AREA      BASE-TEMP       ZONES
APR 1982   1155          0             8                                                   FLAGS BY ZONE
DA  HR PCPN  INT SNOWL  WE  LR TA    MR   RG   ET SMI ROP BFP   SURF  SUBSF BASEF LOWERZ   TOTAL    OBS  1******8
27 240                 709 124                    4.9       0  5.000 13.800 4.000  4.800  27.600  27.550
28 240 0.29 0.19   709 124 7.3 10 0.242 0.30 0.06 5.0  44  81  5.221 16.041 4.185  4.785  30.232  30.200  DDSSSAAA
29 240 0.08 0.05   709 124 7.3  9 0.303 0.13 0.06 4.9  44  80  3.600 15.230 4.365  4.770  27.965  28.800  DDSSSAAA
30 240    0    0   709 123 7.3 12 0.245 0.27 0.08 5.0  44  81  3.393 13.776 4.540  4.755  26.465  27.400  DDDSSSSS
                              V O L U M E  -  C E N T I M E T E R S
      0.37                             0.71                    0.09         0.1           0.63
           0.25                                 0.2                  0.34          0.11           0.65
MAY 1982
 1 240 0.00 0.00   709 123 7.3 12 0.254 0.28 0.08 5.0  44  81  4.115 14.365 4.726  4.741  27.946  28.250  DDDSSSSS
 2 240 1.14 0.30   709 123 7.3 12 0.263 0.40 0.08 5.1  45  80  5.232 16.246 4.934  4.727  31.139  28.150  DDDAAAAA
 3 240 0.70 0.07   709 124 7.3  9 0.311 0.13 0.06 5.1  45  78  4.034 15.994 5.144  4.713  29.885  25.750  DAAAAAAA
 4 240 0.43 0.05   709 124 7.3  9 0.311 0.12 0.06 5.1  45  81  2.360 12.629 5.324  4.699  25.013  24.550  DAAAAAAA
 5 240 0.00 0.00   709 124 7.3 10 0.240 0.18 0.07 5.1  45  83  2.175 10.479 5.487  4.685  22.826  24.900  DDCSSSSS
 6 240 1.41 0.11   709 124 7.3 12 0.247 0.50 0.07 5.2  45  83  4.281 12.318 5.684  4.672  26.956  31.050  DDDAAAAA
 7 240 1.13 0.07   709 125 7.3 12 0.192 0.54 0.08 5.4  46  77  7.222 18.154 5.944  4.659  35.979  36.550  DDDCAAAA
 8 240 0.00 0.00   709 124 7.3 13 0.202 0.39 0.09 5.4  47  74  7.825 22.853 6.226  4.647  41.550  37.000  DDDCSSSS
 9 240 0.00 0.00   709 124 7.3 13 0.211 0.37 0.08 5.4  47  75  7.084 24.008 6.503  4.635  42.230  40.450  DDDCSSSS
10 240 0.00 0.00   709 123 7.3 14 0.180 0.53 0.11 5.5  48  76  8.098 25.551 6.791  4.623  45.063  45.400  DDDQCSSS
11 240 0.01 0.01   709 122 7.3 16 0.164 0.66 0.13 5.6  48  74 11.069 29.932 7.111  4.612  52.723  51.250  DDDQCCAA
12 240 0.70 0.32   709 122 7.3 15 0.167 0.68 0.12 5.6  48  72 13.392 35.360 7.464  4.601  60.816  56.650  DDDQQCAA
13 240 0.14 0.08   709 121 7.3 16 0.176 0.68 0.13 5.7  49  71 14.317 39.818 7.835  4.591  66.562  63.500  LDDQQCAA
14 240 0.03 0.02   709 120 7.3 18 0.175 0.69 0.16 5.7  49  70 14.325 42.876 8.221  4.581  70.004  74.200  LDDQQCCA
15 240 0.11 0.08  1050 119 7.3 19 0.198 0.79 0.19 5.8  50  69 15.741 46.186 8.624  4.572  75.123  82.300  DDDQQCCA
16 240 0.10 0.06  1050 117 7.3 21 0.205 1.07 0.23 6.0  50  69 21.152 52.401 9.069  4.564  87.186  92.250   DDDQQQC
17 240 1.09 0.33  1050 116 7.3 21 0.230 1.39 0.24 6.3  51  66 29.731 63.251 9.594  4.556 107.132 109.000   DDDDQQQ
                               V O L U M E  -  C E N T I M E T E R S
      6.99                             9.39                    1.29         0.86          6.34
           1.50                                 1.97                 3.61          0.59           6.37
```

EXPLANATION OF CODES

DA	Day	**SURF**	Surface flowrate, cms
HR	Hour	**SUBSF**	Subsurface flowrate, cms
PCPN	Precipitation, cm	**BASF**	Baseflow flowrate, cms
INT	Interception, cm	**LOWERZ**	Lower zone flowrate, cms
SNOWL	Elevation of snowline, meters	**TOTAL**	Total computed discharge, cms
WE	Snow water equivalent, cm	**OBS**	Observed discharge, cms
LR	Lapse rate, degrees C / 1000 m	**FLAGS**	Indicators of snow activity on each elevation band
TA	Air temperature at sea level, degrees C	**D**	Dry weather melt occurring
MR	Melt rate, cm/degrees C-day	**R**	Rain melt occurring
RG	Runoff generated, melt+precip-int-soil loss	**S**	Snow on band, no accumulation nor melt
ET	Evapotranspiration, cm/day	**A**	Snow being accumulated
SMI	Soil moisture index, cm	**L**	Dry melt restricted by band transition
ROP	Computed runoff percent	**Q**	Melt, but no RO because of liquid water deficiency
BFP	Computed baseflow percent	**C**	Melt, but no RO because of cold content

Figure 8-2. Example of elevation band output during snowmelt

accommodated to a large degree in satellite observations through the skillful use of image processing. In both types of observations, a heavy forest cover can also obscure the snowline. Again, this can be at least partially accounted for by experienced observers and skilled use of processing techniques.

(3) When an observation of snow cover is obtained, it is compared with the model's current calculation of snow cover or snowline elevation, whether it be the lumped or distributed models described above. A significant difference would suggest a change in the model, particularly if it is confirmed by other indicators described below.

b. Snow-water equivalent. A second field indicator used to verify forecast models is SWE data from snow courses or snow pillows. If automated reporting is available, such as through the SCS SNOTEL network, these data are readily available for

operational use. When used in conjunction with a distributed model, they can help to determine the current SWE state being computed by an element in the model. These measurements can also be used to help estimate the snowline in a basin. An example of a simple approach in using SWE data is shown in Figure 8-3. Here, historical SWE readings for a specific date have been correlated with computed-model SWE for a given elevation band in the basin. The model data are taken from simulation runs made for the basin. Several bands were checked against the observed data to see which had the best correlation and which would be most useful in adjusting the model. Once the model is in the forecast mode, the real-time data are compared against this correlation. If outliers are found, the model should probably be adjusted. The National Weather Service has developed a sophisticated technique using a geographical information system and optimal interpolation (kriging) in which mean areal SWE is calculated from snow measurements. These values are used to update model-simulated SWE via Kalman filtering (Day, Schaake, and Ellis 1989; Day 1990).

c. Streamflow. The final means of checking a forecast model's computation of snowmelt is to compare computed discharge against streamflow observations. Although not necessarily a sensitive indicator in the early stages of snowmelt runoff, this comparison becomes very important in confirming residual SWE volumes carried by the model in late-season melt. This check is viable for both the lumped and distributed models described above. Two different measures of performance are possible: How well does the model compute recently observed streamflow? How reasonable are the streamflow volumes generated by the model when run through the normal recession period? The latter check involves comparing with historical statistics or plots for the period being sampled.

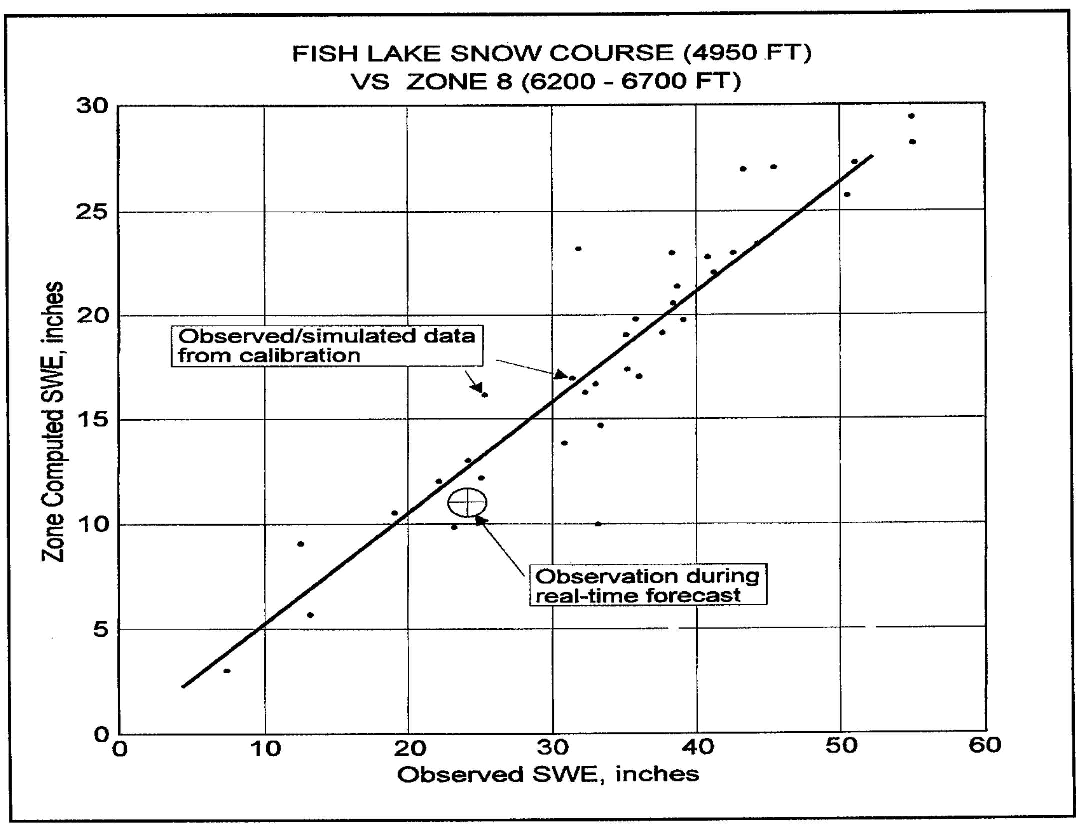

Figure 8-3. Example of correlation—snow band SWE versus snow pillow SWE

Chapter 9
Statistical Analyses

9-1. Overview

Statistical methods are most frequently applied in snow hydrology for water supply forecasting, where measurements of snow and other variables are used to predict spring-summer snowmelt runoff. Because of this widespread practice, the topic will be covered in this chapter as a somewhat special application that is unique to snow hydrology. Aside from this application, statistical methods are also used in other aspects of snow hydrology in much the same way they are used in general hydrology. Frequency methods are used for determining extreme values of SWE or other parameters for design flood or low-flow analyses; multiple regression is employed for regional analyses of various kinds; and stochastic methodologies are used in long-term forecasting. Because they are not especially unique to snow hydrology, however, they will not be described in this manual except in passing. The general principles for applying statistical techniques in hydrology practice are covered in EM 1110-2-1415.

9-2. Data Analysis

Analyzing data for statistical as well as conceptual modeling in snow hydrology requires additional considerations owing to the nature of the environment and data involved. Some factors involved are as follows:

a. Snow data sampling is sometimes not consistent over a period of record. Sampling techniques have changed (e.g., manual to snow pillow) and station sites are sometimes moved.

b. Snow data often have relatively short periods of record compared with precipitation data.

c. Precipitation monitoring is more difficult in a mountainous environment involving snow. Unattended stations, which are subject to capping, are often used and are generally less accurate in measuring short-duration incremental changes.

d. Estimating SWE from precipitation stations may be feasible for high-elevation areas, but it is questionable for areas subject to rain during the winter.

e. Orographic effects and sparse gauge density make it difficult to estimate missing data or area-mean quantities.

Given the above, extra care should be taken in preparing data for use. Double-mass analysis is recommended to evaluate the consistency of the record, and visual or computer screening should be employed to check temporal consistency. It is a common practice to correct older snow-course data to be consistent with more recent snow-pillow records using correlation, if a sufficient overlapping record exists. Cumulative precipitation data sometimes can be used to supplement or simulate high-elevation snow data if a station-to-station correlation exists.

9-3. Frequency Analysis

Frequency analysis in a snow environment is likely to be done on precipitation, SWE, and, perhaps, temperature records. The PMF study described in Chapter 10 employed an extensive meteorological analysis that used depth-duration frequency curves for numerous precipitation stations in the basin being analyzed. Normal annual precipitation (NAP) maps were employed to convert station frequencies to areal representations. Precipitation analysis procedures are described in EM 1110-2-1415. Frequency analysis of SWE data should generally employ the same procedures as those used for precipitation data.

9-4. Water-Supply Forecasting

Water-supply forecasting is the long-term prediction of runoff volume of a specified duration. This term comes from the practice, originating in the 1930s in the western United States, of sampling the winter accumulation of snow to provide an index of runoff in the succeeding spring. Over the years, this basic methodology has evolved into an important and widespread practice that is used for crop management, irrigation planning, flood warning, and reservoir operations. An extensive network of automated snow-monitoring stations, called SNOTEL, have been set up

by the Natural Resource Conservation Service (NRCS) for this purpose, and many agencies process data and coordinate forecasts, such as the NWS, NRCS, USACE, and Bureau of Reclamation, as well as numerous State agencies and electrical utilities. In the West, the NWS and NRCS publish forecasts for over 600 points that appear in the Basin Outlook Reports published by the NRCS and in the Water Supply Outlook for the Western United States, issued jointly by the NWS and NRCS. In California, forecasts are prepared by the State Department of Water Resources, and in the Northeast, NWS publishes water-supply forecasts for the public.

a. Forecasts are usually expressed in terms of a volume of runoff during the months that have operational importance, i.e., April through August, March through July, etc. Winter runoff can also be included to produce January through July forecasts that are important in hydroelectric operations in the Northwest. Traditionally, forecasts have been produced once each month, beginning in January, immediately following measurements of snow and precipitation made on or near the first of the month. In recent years, more frequent forecasting has been made possible by automated hydromet systems.

b. Water-supply forecasts have typically used classic multiple linear regression techniques that incorporate two to five independent variables, as described below. An alternative to the use of multiple regression has been instituted by several agencies in recent years and shows promise as a viable technique for long-range volumetric forecasting. Termed Extended Streamflow Prediction (ESP) by the NWS, this methodology employs continuous simulation models to generate alternative streamflow time series, each reflecting the current state of the basin's condition (snowpack, soil moisture, etc.), combined with future weather conditions from a given historical year. The resulting traces of possible future alternative streamflow are processed as a data sample for statistical analysis. Chapter 10 has further discussion of this approach.

9-5. Multiple Regression Forecast Models

a. Basic equation. The multiple linear regression approach in water supply forecasting uses an equation of the form:

$$Y = a + b_1BF + b_2FP + b_3WP + b_4S + b_5SP \quad (9\text{-}1)$$

where

Y = seasonal streamflow volume

a = regression intercept

b_i = regression coefficients

BF = base-flow index

FP = fall-precipitation index

WP = winter-precipitation index

S = snow-water-equivalent index

SP = spring-precipitation index

(1) The base-flow index is usually the streamflow volume during the fall or early winter, e.g., October-December or November-January. The fall-precipitation index is a sum or weighted sum of monthly precipitation at one or more sites for the fall, e.g., September-November or October-December. The fall-precipitation index and the base-flow index are surrogates for soil moisture. The winter-precipitation index is the cumulative precipitation recorded for that season, say November-March. The snow index is a sum or weighted sum of SWE at several sites for the month usually having the maximum snow accumulation for the season; this is typically April, although it can be March or May. The spring-precipitation index is the same as the winter-precipitation index except for the spring period, e.g., April-June. Not all procedures necessarily use all the variables described above, but as a minimum, winter-snow and precipitation indexes,

a spring-precipitation index, and a fall-soil-moisture index are usually employed.

(2) In some areas of the northwest and southwest United States, another independent variable, the Southern Oscillation Index (SOI), improves the performance of water supply forecasting equations, especially in the early winter before the majority of snow has accumulated. The SOI, an indicator of the El Nino phenomenon, has been shown to be a moderate but significant predictor of winter precipitation and snowpack, with a lead time of as much as 6 months (Koch and Redmond 1991). Historical records of SOI are available, and the index is reported in a timely enough way to be usable in an operational setting.

b. Regression model development. Equation development traditionally begins with an analysis of the station data, employing judgment as to the whether the station should be included in the equation and what the station weighting should be (Hanneford 1993). Such factors as the station's degree of independent correlation with runoff, its location and elevation, and the consistency and viability of past and future data reporting are considered. The station's data are weighted to establish its relative influence in the equation. Remember, in this process relative weighting already exists by virtue of each station's natural mean and variance. A stepwise regression program is then used to select predictor variables and compute the regression equation. At least 15 years of data are necessary for reasonable forecast accuracy. Figure 9-1 is an example of a forecast procedure, laid out in a form that is used operationally in preparing it.

(1) An alternative to the stepwise method of equation development is employing principal components regression. This technique, described by Garen (1992), is used to eliminate aggregating weighted data observations into indices, to address the technical problem of variable intercorrelation, and to more rigorously establish an optimal solution for a given set of data. With it, the independent data are restructured into a equal number of uncorrelated variables via a linear transformation. Each new variable (principal component) is a different linear combination of all the original variables. The new variables are regressed, and variables that have the greatest influence in explaining the variance are selected. These can then be inverted, so that the coefficients are expressed in terms of the original variables. If there was a high degree of inter-correlation in the original data set, this method will result in fewer variables, thereby reducing the loss in degrees of freedom.

(2) With the principal components method eliminating the subjective selection and grouping of data stations for independent variables, a more systematic way of finding the near-optimal combination of predictor variables becomes feasible. Since the number of possible combinations is immense, a computer optimization procedure is required. Garen (1992) has developed an iterative algorithm that appears to be practical and is effective in identifying the strongest variable and constructing a near-optimal model.

(3) One fundamental consideration in developing a multiple regression forecast equation is how to handle precipitation and snowfall that occur after the date of the forecast. Two alternative methods have been used:

- Develop one equation for the season after all data are known; then, at the time of the forecast, use a median or average value to estimate that part of the input that is yet to occur.

- Develop separate equations for each forecast (usually one per month), using only the data known up to that point.

The former method has the advantage of greater stability from month to month and perhaps an advantage in allowing intuitive judgment of the effects of precipitation being above or below normal. However, it has been shown (Garen 1992) that a loss in accuracy results from this method and that it is less rigorous statistically than the second alternative of using separate equations.

9-6. Assessment of Regression Model Accuracy

Once a multiple regression model has been developed, it is necessary to evaluate its ability to represent the

COMPUTATION FORM

FORECASTING RUNOFF FROM LIBBY LOCAL SUBAREA

YEAR__________

Apr - Aug Runoff in Inches = 0.070 (WP) + 0.205 (SP) + 0.047 (SWE) + 0.710 (FRO) - 4.794

					1 Jan	1 Feb	1 Mar	1 Apr	1 May	1 Jun
1.	Forecast Date		Oct	Nov	Dec	Jan	Feb	Mar	Apr	May
2.	FALL RUNOFF (FRO)									
3.	Observed Runoff, Inches—(Observed Libby Inflow - Ft. Steele Observed)									
4.	Sum October + November Runoff, Inches									
5.	Line 4 X 0.710									
6.	WINTER PRECIPITATION (WP)	Weight								
7.	Elko, B.C.	1.00								
8.	Fernie, B.C.	1.00								
9.	Fortine 1 N, MT.	1.00								
10.	Libby R.S. 1 NE, MT.	1.00								
11.	Bonners Ferry 1 SW, ID.	1.00								
12.	Polebridge, MT.	1.00								
13.	Sum Precipitation by Month (Also Equals Sum of Weighted Precip.)									
14.	Sum Precipitation 1 Oct for Forecast Date									
15.	Normal Subsequent Precipitation				43.89	23.58	10.01	0		
16.	Sum Winter Precipitation (Oct thru Apr)									
17.	Line 16 X 0.070									
18.	SNOW WATER EQUIVALENT (SWE)	Weight								
19.	Sullivan Mine, B.C.	1.00								
20.	New Fernie, B.C.	1.00								
21.	Red Mtn., MT.	1.00								
22.	Kimberly, B.C.	1.00								
23.	Weasel Divide, MT.	1.00								
24.	Morrissey Ridge, B.C.	0.50								
25.	Sum of Weighted SWE by Month									
26.	Normal Subsequent SWE					32.58	12.26	0		
27.	Sum (Equals 1 Apr SWE) For 1 Jan Only, SWE = 1.191 X Line 16									
28.	Line 27 X 0.047									
29.	SPRING PRECIPITATION (SP)	Weight (Apr & May)								
30.	Fortine 1 N, MT.	1.00								
31.	Porthill, ID	1.00								
32.	Kaslo, B.C.	1.00								
33.	Whitefish 5 NW, MT.	1.00								
34.	Sum Spring Precipitation by month									
35.	Accumulated Sum Spring Precipitation									
36.	Normal Subsequent Precipitation (Weighted)				24.87	24.87	24.87	24.87	19.23	11.68
37.	Sum				24.87	24.87	24.87	24.87		
38.	Line 37 X 0.205				5.098	5.098	5.098	5.098		
39.	EQUATION CONSTANT				-4.794	-4.794	-4.794	-4.794	-4.794	-4.794
40.	Forecast Apr-Aug Runoff, Inches (Sum of Lines 5, 17, 28, 38, and 39)									
41.	Forecast Apr-Aug Runoff KAF = Line 40 X 251.732									

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Figure 9-1. Water-supply forecast procedure using multiple linear regression

observed data and to assess its accuracy for use as a forecasting tool. This requires an understanding of how to interpret and use error statistics properly, both as a forecast procedure is being developed, and also as it is being executed in a forecasting situation. An in-depth discussion of this subject is beyond the scope of this manual, but a summary discussion of several analysis methods that are often used in practice will be presented. There is generally no lack of error statistics available for the analyst who is using modern statistical computer programs. The problem in practice generally lies in understanding what the error factor means, and in applying it meaningfully in forecasting or analysis. Further discussion on this topic can be found in EM 1110-2-1415, as well as in numerous textbooks and manuals.

a. Evaluation criteria. There are several criteria that are commonly used to evaluate multiple regression models (McCuen 1985):

(1) Rationality of the coefficients.

(2) Relative importance of the predictor variables.

(3) Characteristics of the residuals.

(4) Coefficient of multiple determination.

(5) Standard error of the estimate.

Coefficient rationality is determined by subjective inspection, by substituting possible values for variables and noting the results in the dependent variable. Basic checks might include the following:

(1) Is the change in forecast logical when a predictor variable is changed in a certain direction?

(2) Is the forecast reasonable when variable extremes are encountered?

A further check of rationality is to examine the relative importance of the predictor variables. This may be subjectively evaluated as above, or analytical procedures can be used. If a certain variable is of little consequence in determining the prediction, it might be deleted from the equation for simplification. On the other hand, if the variable should be more influential than it is showing, then a reexamination of the model is necessary.

b. Analysis of residuals. In correlation analysis, the residual is the unexplained difference between the predicted and observed value of the independent variable, as illustrated in Figure 9-2. By definition, through the application of the least squares objective function, the sum of the residuals must equal zero. However, this does not guarantee that the model is not biased. If, for instance, the residuals tend to be positive for low values of X but negative for high values of X, then bias exists and a nonlinear model may need to be used. Plots of residuals can be made in various ways to check the validity of the model. A plot of residuals as a function of the dependent variables would display the bias as a function of observation magnitude, while a probability plot of the residuals might help verify the assumption that they are normally distributed in the Y direction.

c. Coefficient of multiple determination (R^2). The coefficient of multiple determination is the proportion of the variance of the dependent variable that is explained by the regression equation. A coefficient of determination of 0.25 means that 25 percent of the variance of the Y variable about its mean is accounted for and 75 percent is not explained by the regression equation. The range of R^2 is between 0 and 1.0, with the value of 0 indicating that Y is not related to any of the predictor variables. In general, this statistic provides a relative measure of the accuracy of the equation in making future predictions—assuming, of course, that the data sample is representative of the total population.

d. Standard error of estimate. The standard error of estimate is the standard deviation of the residuals, computed as the square root of the sum of the squares of the errors divided by the degrees of freedom (df). By definition, assuming a normal distribution of the residuals, two-thirds of the estimates will fall within plus or minus one standard error; 16 percent will be above the mean plus one standard error, and 16 percent will be fall below the mean minus one standard error. This is illustrated in Figure 9-2. Other

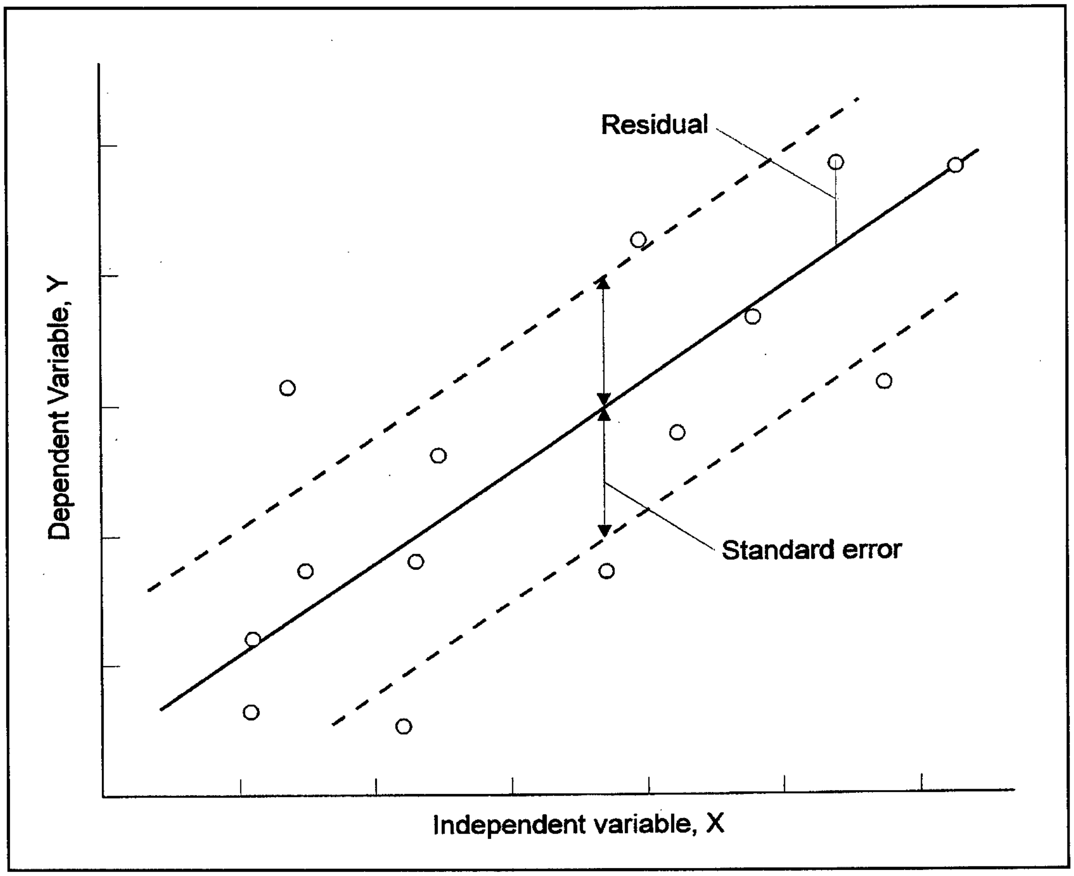

Figure 9-2. Correlation error analysis

cumulative normal curve found in statistical reference books.

(1) A problem frequently encountered in water-supply-forecasting practice is properly accounting for the value of degrees of freedom. The value of the degrees of freedom is obtained by subtracting the number of variables (independent and dependent) from the number of data points (years) defining the relationship. It is common practice to use a df equal to the number of major variables—snow, precipitation, etc. Yet, these variables may in fact have been made up of a number of stations that have been subjectively selected and weighted. In reality, the loss of degrees of freedom may be higher than the number of nominal variables contained in the equation, and a plot such as Figure 9-2 may be optimistically portraying the ability of this forecast to perform in the "real world" of actual future forecasts. This has been borne out in general by comparing with the cross-validation technique described below.

(2) In recent years, a more realistic portrayal of forecast accuracy in an actual forecasting situation has been obtained by using the cross-validation or "jack-knife" procedure. Here, one observation is removed from the data set, and the regression coefficients are calculated. These coefficients are used to predict the dependent variable for the withheld observation. The withheld data are returned and the next observation is removed. This process is repeated until a "forecast" has been made for all of the observations, using coefficients that do not reflect that data. A standard error is then calculated from these "forecasts." Comparison of error estimates using this method with traditionally computed standard errors shows that the traditional errors tend to underestimate the more rigorously computed standard errors.

Chapter 10
Snowmelt Runoff Analysis for Engineering and Forecasting Applications

10-1. Problem Definition, Selection of Methodology

a. General. This chapter will discuss the practical aspects of analyzing snowmelt runoff for specific applications normally encountered within USACE. Discussed are the considerations needed in deciding on the methodology to use, the degree of detail with which snowmelt is to be analyzed, the selection of the modeling approach that should be used, and specifics of the analysis and simulation for specific applications. EM 1110-2-1417, Flood Runoff Analysis, contains a discussion of developing a hydrological engineering investigation in concert with the stage of planning and design.

b. Overview of applications and approaches. There are numerous alternatives for determining the best approach for computing snowmelt in hydrologic engineering analysis and forecasting. These range from simplified assumptions on discrete storm events to detailed simulation using energy budget principles and a distributed definition of the watershed. The choice depends on the degree of detail called for, the degree to which snow is a factor in affecting runoff, the resources available to do the analysis or maintain operational-forecasting capability, and data availability. For applications involving snowmelt, the choice for analysis is complicated by the need to consider a more detailed basin definition than for rain only, and by the range of options to consider in computing snowmelt. Table 10-1 summarizes some possible analysis alternatives and how they relate to given types of applications.

c. Selection of models. Chapter 11 contains summary guidance that will help with the selection of hydrologic models currently available for use in analysis and forecasting, and Appendix F presents detailed descriptions of the computer models. It is well to remember that successful application of a model depends upon the skill and knowledge of the user and a thorough understanding of the physical processes involved.

Table 10-1
Snowmelt Options[1]

		Basin Configuration		Melt Calculation			
Application	Example	Lumped	Distributed	Snow Conditioning	Simplified[2]	Temperature Index	Energy Budget
Single-event analysis-Rain-on-snow	Hypothetical floods in coastal mountains	Yes	Possibly	Assumed "ripe"	Possibly	Possibly	Possibly
Single-event analysis-Snow (plus rain)	Hypothetical floods in interior basins	Yes	Yes	Assumed "ripe"	No	Yes	Yes
Single-event forecasting-Rain-on-snow	Short-term flood forecasting	Yes	Yes	Optional	Possibly[3]	Yes	No
Single-event forecasting-Snow (plus rain)	Short-term flood forecasting	Yes	Yes	Optional	No	Yes	No
Continuous simulation, any environment	Long-term flood and drought forecasting; detailed design analysis	No	Required	Required	No	Yes	Possibly
Detailed simulation in small watersheds	R&D applications; analysis for detailed design; special applications	No	Required	Required	No	No	Yes

[1] Qualitative indicator shown for type of option that might typically be used for application. This is a guideline only. "Yes" or "No" indicates suggested option.
[2] Simplified approach might be to assume a constant- or variable-moisture input due to snowmelt.
[3] Would be appropriate only in situations where snowmelt is small compared with rain.

10-2. Hypothetical Floods

Developing a hypothetical flood entails using a hydrological model of some type to generate a streamflow hydrograph, given rain and snowmelt input of a specified magnitude. Two examples might be floods of estimated frequency for an ungauged area, using rainstorms of specified frequency for input, and an inflow design flood (IDF) for a proposed or existing dam, using probable maximum precipitation (PMP) as input. If snow is involved, then the decision must be made as to how snowmelt runoff is best computed, given the application being used and the range of alternative methodologies summarized on Table 10-1. In the following paragraphs, some alternative methods with varying complexity are described and two examples are given.

a. Simple approaches. In certain situations, a simple method for snowmelt runoff may be entirely satisfactory or, in fact, be required. A basin with rain on snow, in which rainfall is the dominant source of runoff during a flood, would not need snowmelt to be computed with a lot of detail, particularly in early stages of project planning. At its simplest, an assumed fixed rate of melt could be added to rainfall, or a variable rate could be estimated independently on the basis of a temperature-index approach. The snow could be considered fully primed prior to the onset of rain, and an adequate initial amount of SWE could be assumed available to contribute fully to the flood peak. These assumptions should be verified with an investigation of historical flood patterns and perhaps some sensitivity testing.

b. Example of a 100-year flood derivation, event-type model. The following is a hypothetical problem that uses a lumped-event model to derive a design flood. In this example, the temperature-index method is used to compute snowmelt, but the melt-rate factor was carefully estimated using the energy budget equation, and this factor was checked for sensitivity in affecting the outcome.

(1) Setting. This is assumed to be an ungauged watershed in which a synthetic unit hydrograph has been derived. A 100-year flood is to be derived for a reconnaissance study by using a 100-year storm taken from *NOAA Atlas II.* The only data on snow are based on nearby weather records that show that as much as 50.8 cm (20 in.) of snow has accumulated in midwinter. An average snow depth of 45.7 cm (18 in.) is assumed for the basin as an average. With an assumed snow density of 20 percent, this yields an initial SWE of 9.1 cm (3.6 in.). Table 10-2 is a summary of the initial assumptions for this problem.

Table 10-2
Summary of Input for Design Flood Derivation, Simple Approach

Item	Description
Drainage area	75 km (29 miles^2)
Forest cover	25 percent
Snyder's IUG[1] coefficients	$T_p = 2.1$; $C_p = 0.40$
Computation interval	1 hr
24-hr precipitation	9.4 cm (3.7 in.)
Maximum hourly precipitation	1 cm (0.4 in.)
Loss rate	Constant: 0.1 cm (0.04 in.)/hr
Initial snow depth	45.7 cm (18 in.) (basin mean)
Initial density	20 percent
Computed initial SWE	9.1 cm (3.6 in.)
Maximum air temperature	12 °C (54 °F) mid elev of basin
Snow condition	Assumed ripe

[1] IUG = Instantaneous Unit Graph.

(2) Melt determination. For this derivation, the temperature index approach will be used in computing hourly snowmelt. Since the basin is relatively open and subject to high-condensation melt, the melt-rate coefficient must be chosen carefully. This is done using Equation 5-19. With T_a = 12 °C (54 °F), v = 24 km/hr (15 mph), P_r = 8.9 cm (3.5 in.), and k = 0.7, the 24-hr snowmelt would be about 8.1 cm (3.2 in.). This suggests a value for C_m of 0.13 to 0.16 in Equation 6-1, using a base temperature of 0 °C (32 °F). A coefficient of 0.14 will be used initially and a sensitivity test done to see its relative influence. A temperature sequence for the storm will begin at near freezing and increase to the maximum in time to produce maximum melt that contributes to the flood peak.

(3) Model output. HEC-1 was used to simulate the conditions described above. Figure 10-1 is a listing of the output. A peak flow of 175.6 cu m/s (6200 cfs) results from the conditions assumed. Figure 10-2 is a plot of the hydrograph.

(4) Analysis of results. Several simulations were made with varying melt-rate factors. The results are shown on Figure 10-3. An incremental change in C_m by 0.02 results in about a 5- to 6-percent change in the peak of the design flood. The assumed melt-rate coefficient of 0.14 seemed reasonable for the physical conditions involved and for the design flood magnitude being derived. The initial SWE assumption of 9.1 cm (3.6 in.) was verified by inspection. There was approximately 4.8 cm (1.9 in.) of snowmelt before the maximum moisture input to the flood, indicating that the SWE could be reduced by 60 percent and still be fully contributing to the peak of the flood.

c. Detailed analyses. A more thorough analysis than discussed above would be required for detailed design studies and certain operational studies. Elements that would be required in a detailed study that are not reflected in the above example could include the following.

(1) Distributed modeling. This is generally used for rain-on-snow situations. For some spring-summer snowmelt areas, where summer rainfall is not highly significant, it may be possible to use a snow cover depletion curve as described in Chapter 8.

(2) Use of energy budget equations. If snowmelt is significant in influencing the magnitude of the flood peak, then energy budget equations should be used for computing it. This is necessitated by the need to better quantify the melt-rate magnitude as a function of the physical elements involved.

(3) Continuous simulation modeling. For settings requiring lengthy periods of simulation (e.g., spring-summer snowmelt), evapotranspiration and other factors should be taken into account.

(4) Model calibration. The problem with calibration using energy budget equations is finding the necessary solar radiation, wind, dew point, and temperature data that are required. A partial solution would be to employ the temperature-index methodology to calibrate the soil-moisture accounting and runoff-transformation portion of the model, using an extended period of record. The energy budget factors could then be calibrated on a shorter period of record or for a portion of the basin for the more difficult to obtain data. This would require a computer model that has the option of using both a temperature and energy budget approach in computing snowmelt.

(5) Thorough analysis of initial snowpack conditions. Where snowmelt volume is a dominant factor in determining the magnitude of the design flood, the initial size of the snowpack must be carefully derived. This implies using an independent statistical analysis of historical data, a special hydrometeorological analysis for extreme flood derivations, or continuous simulation during the winter-accumulation season for a period that spans enough years of record to provide a viable statistical sample. In addition to snowpack volume, the horizontal and vertical distributions need to be derived. Snow-condition effects also need to be developed, at least for rain-on-snow conditions. For spring snowmelt flood derivations, a ripe initial snowpack can be assumed since flood simulations typically begin in early spring.

(6) Melt-sequence derivation. The meteorological factors that are used as independent variables for computing melt must be carefully derived on the basis of historical sequences, using a degree of maximization appropriate for the design flood magnitude.

(7) Thorough analysis of rain-on-snow variations. Virtually every climatic region experiences a mixture of rain-on-snow alternatives, be it during the winter where rain dominates or during the springtime when rain may or may not be a significant factor in defining the design flood. The rainstorm magnitude and areal extent must be carefully developed, considering the relative magnitude of the design flood, ensuring that an appropriate combined probability of occurrence is reflected in the snow and rainfall combination.

d. Optimal conditions for probable maximum floods. Following standard USACE guidance, a PMF derivation requires maximization of the flood's components so that the resulting flood runoff is the

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HYDROGRAPH AT STATION 1

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DA	MON HR	ORD	PRECIP	TEMP	SNOMELT	SNOLOSS	SNOEXCS	RAIN	RAINLOS	RAINEXS	SNO+RAIN	LOSS	EXCESS	COMP Q
25	DEC 0000	1	0.00	40.0	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	20
25	DEC 0100	2	0.00	41.0	0.04	0.04	0.01	0.00	0.00	0.00	0.04	0.04	0.01	27
25	DEC 0200	3	0.00	42.0	0.05	0.04	0.01	0.00	0.00	0.00	0.05	0.04	0.01	55
25	DEC 0300	4	0.00	43.0	0.06	0.04	0.02	0.00	0.00	0.00	0.06	0.04	0.02	102
25	DEC 0400	5	0.00	44.0	0.06	0.04	0.03	0.00	0.00	0.00	0.06	0.04	0.03	163
25	DEC 0500	6	0.00	45.0	0.07	0.04	0.03	0.00	0.00	0.00	0.07	0.04	0.03	235
25	DEC 0600	7	0.00	46.0	0.07	0.04	0.04	0.00	0.00	0.00	0.07	0.04	0.04	315
25	DEC 0700	8	0.08	46.7	0.08	0.02	0.06	0.08	0.02	0.06	0.16	0.04	0.12	482
25	DEC 0800	9	0.09	47.3	0.08	0.02	0.06	0.09	0.02	0.07	0.17	0.04	0.13	807
25	DEC 0900	10	0.09	48.0	0.08	0.02	0.07	0.09	0.02	0.07	0.17	0.04	0.14	1180
25	DEC 1000	11	0.10	48.7	0.09	0.02	0.07	0.10	0.02	0.08	0.19	0.04	0.15	1507
25	DEC 1100	12	0.10	49.3	0.09	0.02	0.08	0.10	0.02	0.08	0.19	0.04	0.16	1798
25	DEC 1200	13	0.11	50.0	0.10	0.02	0.08	0.11	0.02	0.09	0.21	0.04	0.17	2061
25	DEC 1300	14	0.11	50.7	0.10	0.02	0.08	0.11	0.02	0.09	0.21	0.04	0.17	2303
25	DEC 1400	15	0.13	51.3	0.10	0.02	0.09	0.13	0.02	0.11	0.23	0.04	0.20	2538
25	DEC 1500	16	0.14	52.0	0.11	0.02	0.09	0.14	0.02	0.12	0.25	0.04	0.21	2789
25	DEC 1600	17	0.22	52.7	0.11	0.01	0.10	0.22	0.02	0.20	0.33	0.04	0.30	3125
25	DEC 1700	18	0.25	53.3	0.12	0.01	0.10	0.25	0.02	0.23	0.37	0.04	0.33	3610
25	DEC 1800	19	0.26	54.0	0.12	0.01	0.11	0.26	0.02	0.24	0.38	0.04	0.34	4160
25	DEC 1900	20	0.31	54.0	0.12	0.01	0.11	0.31	0.03	0.28	0.43	0.04	0.39	4703
25	DEC 2000	21	0.40	54.0	0.12	0.01	0.11	0.40	0.03	0.37	0.52	0.04	0.48	5338
25	DEC 2100	22	0.30	54.0	0.12	0.01	0.11	0.30	0.03	0.27	0.42	0.04	0.38	5967
25	DEC 2200	23	0.16	54.0	0.12	0.02	0.10	0.16	0.02	0.14	0.28	0.04	0.24	6196
25	DEC 2300	24	0.12	54.0	0.12	0.02	0.10	0.12	0.02	0.10	0.24	0.04	0.20	5937
26	DEC 0000	25	0.12	54.0	0.12	0.02	0.10	0.12	0.02	0.10	0.24	0.04	0.20	5507
26	DEC 0100	26	0.11	52.7	0.11	0.02	0.09	0.11	0.02	0.09	0.22	0.04	0.19	5112
26	DEC 0200	27	0.11	51.3	0.10	0.02	0.09	0.11	0.02	0.09	0.21	0.04	0.18	4760
26	DEC 0300	28	0.11	50.0	0.10	0.02	0.08	0.11	0.02	0.09	0.21	0.04	0.17	4443
26	DEC 0400	29	0.10	48.7	0.09	0.02	0.07	0.10	0.02	0.08	0.19	0.04	0.15	4154
26	DEC 0500	30	0.09	47.3	0.08	0.02	0.06	0.09	0.02	0.07	0.17	0.04	0.13	3865
26	DEC 0600	31	0.08	46.0	0.07	0.02	0.06	0.08	0.02	0.06	0.15	0.04	0.12	3565
26	DEC 0700	32	0.00	45.0	0.07	0.04	0.03	0.00	0.00	0.00	0.07	0.04	0.03	3187
26	DEC 0800	33	0.03	44.0	0.06	0.02	0.04	0.03	0.01	0.02	0.09	0.04	0.06	2721
26	DEC 0900	34	0.01	43.0	0.06	0.03	0.02	0.01	0.01	0.00	0.07	0.04	0.03	2292

EXPLANATION OF CODES

DA MON HRMN: DAY, MONTH, HOUR, MINUTE
ORD: ORDINATE NUMBER
PRECIP: PERIOD PRECIPITATION, in
TEMP: PERIOD TEMPERATURE, , degrees F
SNOMELT: COMPUTED PERIOD SNOWMELT, in
SNOLOSS: COMPUTED PERIOD SNOWMELT LOSS, in
SNOEXCS: PERIOD SNOWMELT EXCESS, in
RAIN: BASIN PERIOD RAINFALL, in
RAINLOS: COMPUTED PERIOD RAIN LOSS, in
RAINEXS: COMPUTED PERIOD RAINFALL EXCESS, in
SNO+RAIN: TOTAL OF SNOWMELT PLUS RAINFALL FOR PERIOD, in
LOSS: TOTAL OF SNOW LOSS AND RAIN LOSS FOR PERIOD, in
EXCESS: TOTAL OF SNOW EXCESS AND RAIN EXCESS FOR PERIOD, in
COMP Q: COMPUTED DISCHARGE FOR PERIOD, cfs

Figure 10-1. HEC-1 output

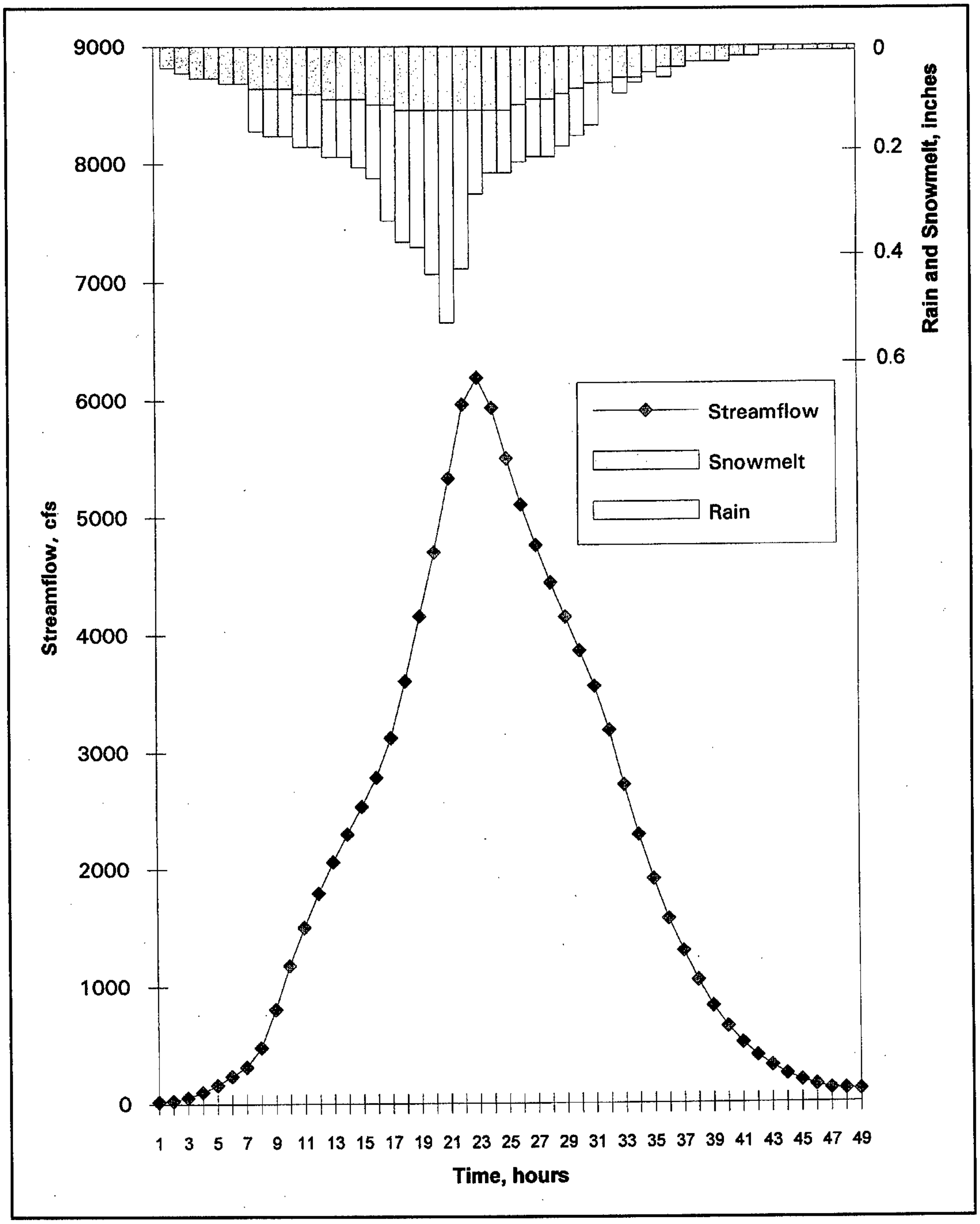

Figure 10-2. Plot of example design flood

maximum reasonably possible for a given basin. For snowmelt regimes, the components discussed below must be examined and maximized. The temperature index cannot be relied upon for a PMF derivation because of the lack of uniformity among basins of different environments, the significant changes in snowmelt rates that may take place within a given basin because of factors other than air temperature, and the danger of extrapolating to conditions beyond the limits to which the index applies.

Cm in/degF	Max Melt in	Peak kcfs
0.08	0.07	5.3
0.10	0.09	5.6
0.12	0.10	5.9
0.14	0.12	6.2
0.16	0.14	6.5
0.18	0.15	6.8
0.20	0.17	7.1

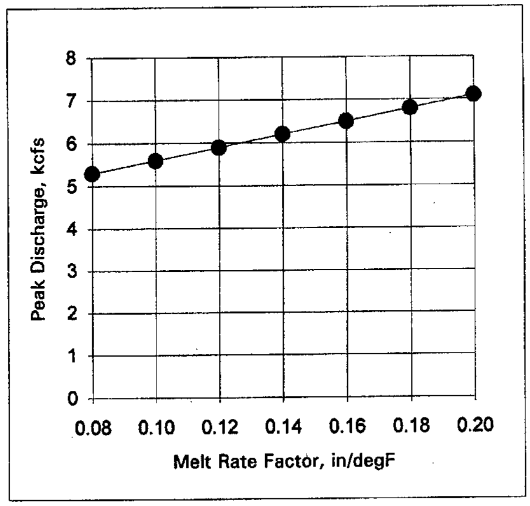

Figure 10-3. Sensitivity of melt-rate factor

(1) Optimal snowpack conditions. For spring-summer PMFs, the maximum possible SWE is usually derived from detailed studies of potential total-winter precipitation. The studies may use derived relationships in which the extreme can be readily inferred and generalized; i.e., maximum winter-season precipitation versus drainage and normal annual precipitation. For rain-on-snow conditions, it is usually assumed that sufficient water equivalent exists to provide snowmelt continuously through the rainstorm. A conservatively high assumption about snow condition is also appropriate; typically, an antecedent storm is assumed, so this would lead to ripe or nearly ripe snowpack conditions for the PMF itself.

(2) Optimal snowmelt conditions. For spring-summer snowmelt floods, the critical flood-producing meteorological conditions are those in which the winter snowpack accumulates with no significant melt, followed by a cold spring with minimum snowmelt and a continued increase in the snowpack. After the maximum snowpack has accumulated, there is a conditioning period during which the melt is moderate; the snowpack and underlying soil are conditioned to produce maximum runoff throughout the basin, and the snow-surface albedo may approach its minimum value. Finally, the meteorological factors affecting snowmelt are allowed to increase to their maximums, at a time when the heat input to the basin can be near its seasonal maximum. The prolonged period of continuous high-heat input is important in producing the maximum flood peak. Then, the runoff rates may approach the snowmelt rates for the snow-covered area, contributing to runoff at the time of the flood peak as an equilibrium inflow-outflow condition.

(a) The meteorological components used in the energy budget equations depend upon the degree of forest cover, as outlined in Chapter 5. The various components must be maximized individually using historical records as a guide. Examples of derived meteorological factors are given in the example below.

(b) For rain-on-snow settings, the temperature and wind-velocity time series during the rainstorm are again determined by considering historical conditions and extrapolating to reasonable maximum characteristic values.

(3) Optimal snow and rain combinations. The PMF derivation needs to have examined alternative possibilities for rain-snow combinations, most likely by simulating alternative scenarios. For spring-summer events, the critical combination is likely to be a large snowpack combined with a maximum melt sequence that is interrupted by a spring rainstorm. However, it may be unreasonable to maximize all these components, so a decision needs to be made

about which factor should be the dominant one in creating the PMF. Bearing on this is whether volume of runoff is a critical factor, as it might be for an IDF for a large reservoir or system of reservoirs.

(a) For instance, a maximized snowpack in conjunction with a severe but not maximized spring rainstorm may produce a flood with lower peak but higher volume than if a lower snowpack with a maximized spring rainfall were used. The former may be more critical for a large storage reservoir, while the latter would be appropriate for projects having less storage. A factor to consider in this analysis is whether the storage can be assumed to be fully available. The standard practice is to assume water supply forecasts will be accurate enough to dictate maximum drawdown prior to the flood—given that a large enough snowpack is involved. However, outlet and downstream channel conditions that might restrict drawdown rates under the generally wet winter conditions that would be associated with the PMF need to be considered.

(b) For rain-on-snow regimes, determining the rain-snow combination is less problematic. With rainfall dominating in governing the flood peak and volume, the SWE magnitude and temperature sequence would not be extrapolated to maximum values, but might still represent a relatively high probability of occurrence.

e. Example of detailed flood derivation. The following example is taken from a PMF study for the Columbia River Basin by the North Pacific Division, with assistance from the Hydrologic Engineering Center (USACE 1969). In this study the SSARR model was used to simulate the design flood for the entire basin at the site of Bonneville Dam (673 395 square kilometers (260 000 square miles)). The flood resulted from a maximized winter accumulation of snow combined with a critical sequence of spring temperatures interrupted by two spring rainstorms. Flood-control storage space was available in upstream storage reservoirs at the beginning of the flood, and the flood was regulated as much as possible by these projects according to a predetermined operating plan. A detailed explanation of the work is given in the 1969 report. Excerpts from that report are shown below as a general illustration of the concepts involved.

(1) Winter snow accumulation. A comprehensive study was undertaken to determine the initial SWE for the snowmelt runoff simulation. Precipitation frequency curves were developed for the October-April period for 54 stations in the basin, and from these, 100-year values were computed. Several approaches were then investigated for determining a relationship between the 100-year depth for subbasin areas as a function of the 100-year depth for the total drainage. An elliptical isopercental pattern for the 7-month precipitation was also derived, as shown on Figure 10-4. Then, using both statistical and hydrometeorological methods, a value representing the total basin PMP was adopted—this was established as 130 percent of the NAP. This value could then be distributed to subbasins using the isopercental pattern and the drainage area-precipitation depth relationship.

(2) Snowmelt calculation. The generalized energy budget equation for snowmelt in partly forested areas (Equation 5-25) was used for all subbasins. This required the derivation of time series for several meteorological variables during the 15 April through 31 July melt period. These variables were obtained by evaluating historical data and by referring to the snow investigations data and relationships.

(a) Examples of derived temperature and dew-point sequences are shown in Figure 10-5. The dew point was assumed to have a -9.4 °C (15 °F) depression from air temperature, except during the spring rainstorms, when this was reduced to -16.7 °C (2 °F). A lapse rate of -15.9 °C (3.3 °F) was used for both of these factors in applying them to different elevations in the basin.

(b) Solar radiation was computed as a sequence of daily averages with no attempt made to evaluate the slight variations with latitude within the basin. Except for the periods of rain and short transition periods, near-maximum values for the location, reflecting cloudless skies, were assumed to prevail. The adopted values of insolation were based on Figure D-8. An assumed albedo pattern decreasing from 80 percent in mid-April to 40 percent in July was derived. The shape of this function is based on snow investigations

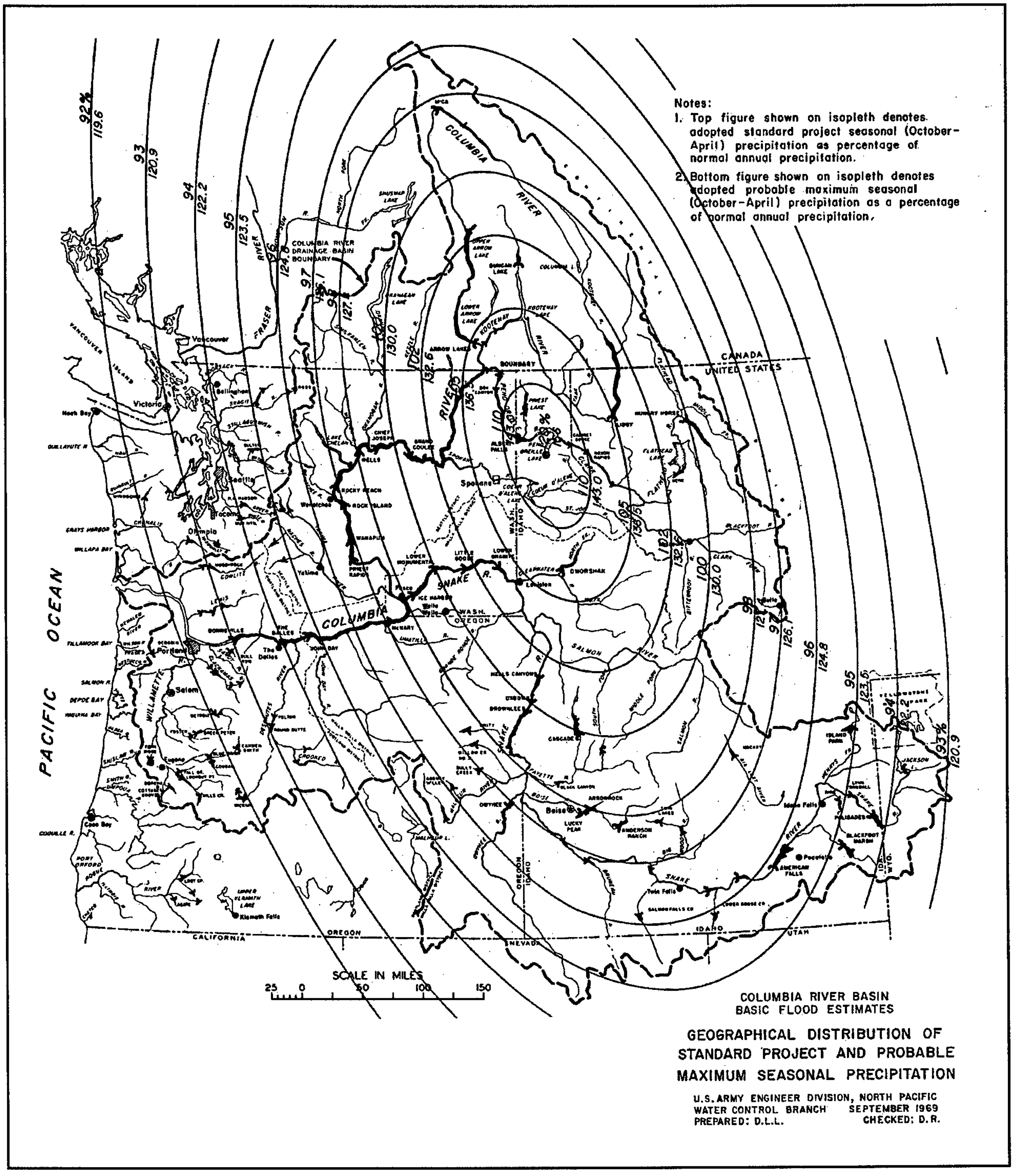

Figure 10-4. Geographical distribution of Columbia River basin PMP

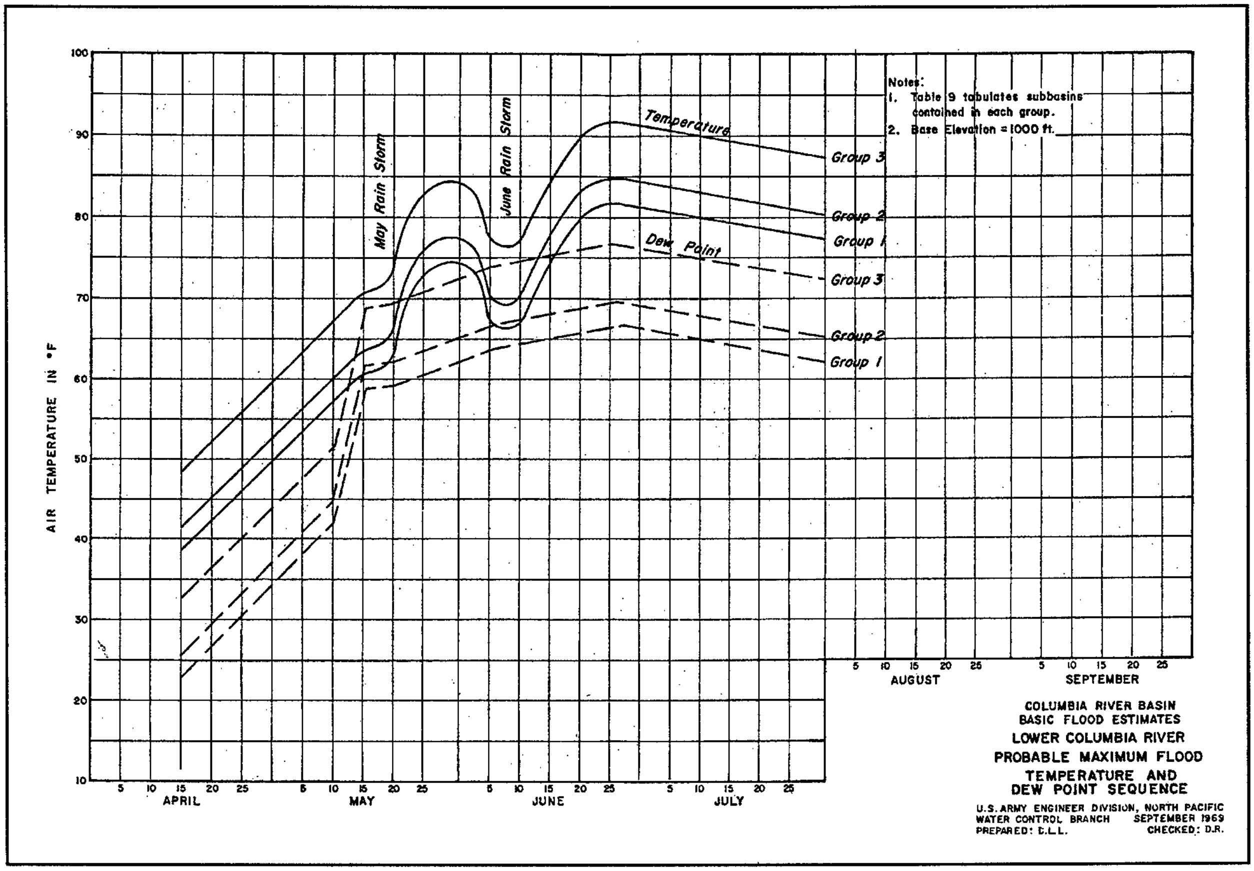

Figure 10-5. Temperature and dew-point sequence

data (Figure 5-5). The insolation and albedo patterns used in the study are shown on Figure 10-6.

(c) Wind velocity was assumed to be 24 km/hr (15 mph) at the 15.2-m (50-ft) level throughout the melt period, increasing to 48.3 km/hr (30 mph) during the two spring rainstorms.

(3) Spring rainstorms. Separate 3-day spring rainstorms were assumed for May and June. The depth for these storms was determined by subtracting October-April (and October-May) seasonal precipitation totals from October-May (and October-June) totals for each of the precipitation stations used in the analysis. These were normalized to percent of NAP for distribution throughout the basin. In effect, the monthly total was assumed to fall in the 3-day period. The two rainstorms are apparent in affecting the other meteorological variables in the above figures.

(4) Basin simulation. The model of the Columbia basin included 63 subbasin watersheds that fed runoff into an extensive river-reservoir simulation model. The river model included the effects of irrigation diversions, lakes, and reservoir operations. The resulting PMF at The Dalles Dam (613 830 square kilometers (DA = 237 000 square miles)) is shown on Figure 10-7.

10-3. Reservoir Regulation Studies

a. Overview. There are a variety of hydrological studies that may be required in support of a reservoir-regulation mission. Flood-control rule curves may need refining; new environmental regulations may require reconsidering of established rule curves; reallocating of storage may be proposed; forecasting procedures may need improving; etc. Such studies have the potential for requiring a relatively

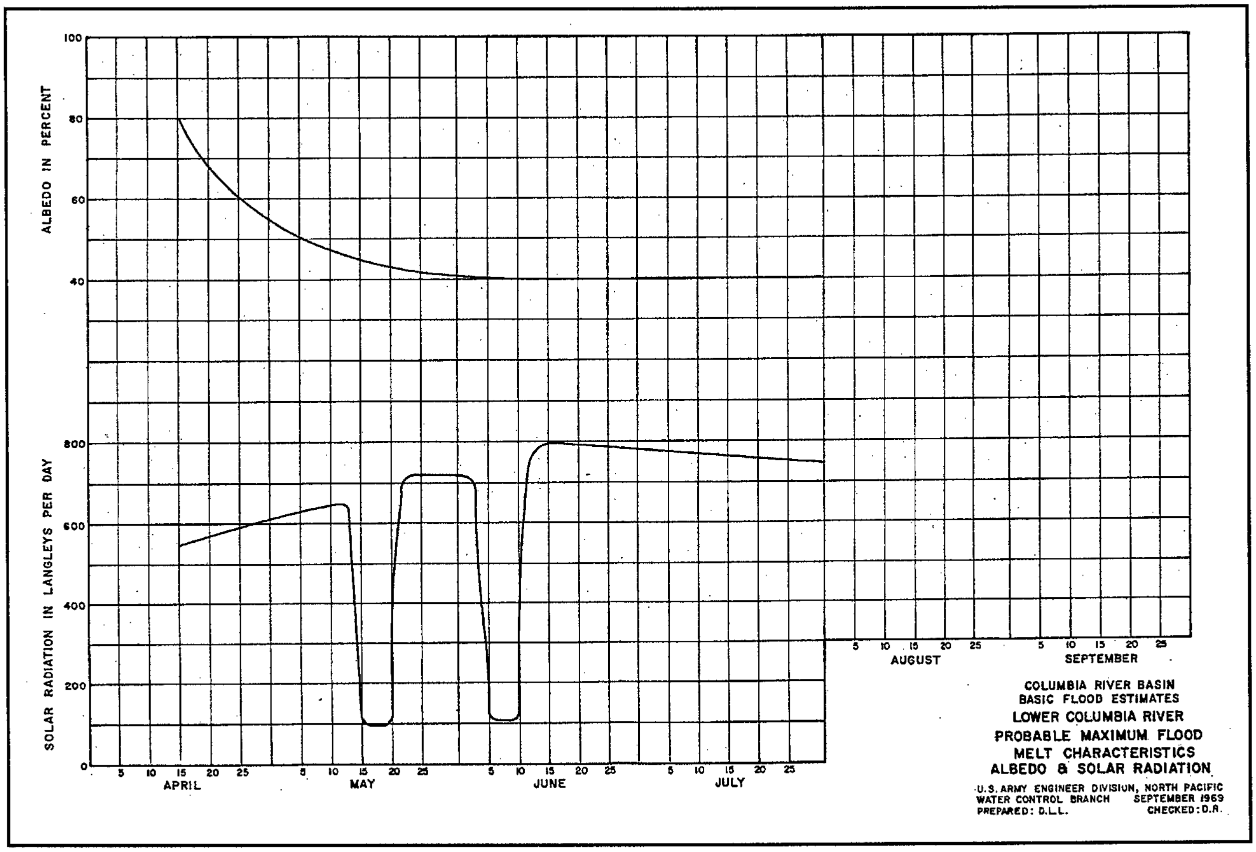

Figure 10-6. Solar radiation and albedo sequences

sophisticated study approach since regulation issues are often complex, involve significant project benefits, and have high public and political visibility. Because water-supply, as well flood-control, considerations may be involved, the use of continuous simulation modeling employing distributed models may be needed. In an environment with snowmelt, the following types of studies may be required.

(1) Water-supply forecasting. Water-supply forecasting procedures described in Chapter 9 may need developing or improving. It is common practice to update statistical procedures periodically to incorporate a larger statistical sample and make necessary corrections. If ESP procedures are to be used as described further in this chapter, continuous modeling is required.

(2) Streamflow forecasting. The development of streamflow forecasting models for guiding reservoir regulation can require extensive model calibration and testing and setting up of a real-time forecasting process if not already existing. The type of model structure would have to be decided upon depending upon the needs and type of snow environment (Chapter 4).

(3) Flood-control curves. Evaluation of flood-control rule curves may require specialized simulation studies that use more complex models for snowmelt runoff. An example of one such study is described below.

(4) Seasonal regulation studies. If operating guidelines are modified in any way, the effects of the changes need to be evaluated. This includes determining downstream flood-frequency curves and reservoir elevation-frequency curves, having the ability to meet desired operating objectives, etc. Typically, such studies use a reservoir system model,

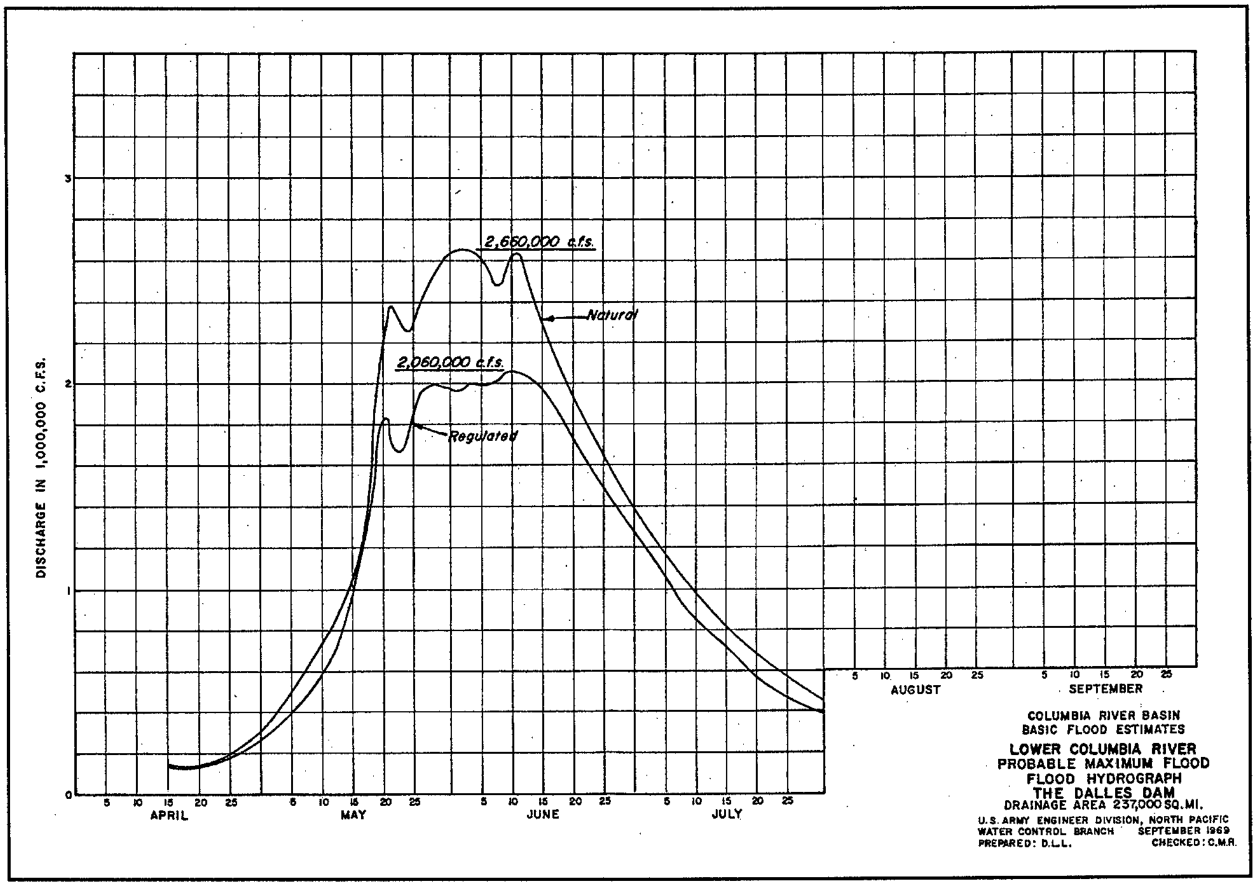

Figure 10-7. PMF, Columbia River at Dalles Dam

perhaps operating on a monthly time step, and using historical observed streamflow, rather than a runoff model. For flood-control evaluations, of course, a short-term computation interval is required. If the evaluation requires using synthetic hydrographs, then a snowmelt runoff model would be required. In reservoir studies for a snow environment, the ability to use water-supply forecasts in guiding reservoir drawdown would normally be assumed; however, a realistic portrayal of forecast error needs to be reflected in the studies. The assessment of this error itself requires a careful analysis.

b. Example of reservoir rule curve study. Snowmelt runoff modeling was employed in a 1987 analysis of rule curves for flood-control reservoirs in the Columbia River basin. In this area flood-control drawdown is based primarily upon water-supply forecasts using flood-control rule curves. These curves, however, include a factor of safety to account for unforecastable spring rainfall. The problem was to evaluate the magnitude of this factor of safety for all ranges of snow and rainfall magnitudes. There is limited historical experience of rain-on-snow events; several have happened, but in conjunction with larger snowpacks. Needed in this study was an evaluation of the effect of rain falling primarily on low snowpacks to ensure adequate flood control in low-snow conditions. This required the development of synthetic rain-on-snow combinations.

(1) For this analysis, a distributed (elevation-band) model, operated continuously through the year, was used. It was calibrated on the period of record, in most cases, using the temperature index for computing snowmelt. Several selected years, representing a range of snow-accumulation magnitudes, were used for the analysis, with emphasis placed on the

low-snow events. In a separate analysis, spring rainstorms were examined for depth, duration, and timing. Storms of specific frequency (100-year storms were used primarily) were derived using several different historical timing patterns. The synthetic floods were then created by simulating the known snowmelt situation with the several alternatives of possible 100-year spring rainfall imposed. Figure 10-8 is an example of four floods so derived, showing the historical reservoir inflow for a relatively low snowmelt year (1973) plotted against the synthetic floods.

(2) With knowledge of the potential reservoir inflow resulting from the spring rainfall, rule curves could be objectively established to make sure that storage space was available to contend with the spring rainfall and not change the overall downstream flood-control capability. This study resulted in a reduction in the flood-control requirement at several reservoirs for low snowpack conditions, which benefited operations for other project purposes. The existing and proposed flood-control rule curves are shown in Figure-10-9.

10-4. Operational Forecasting

a. Overview. Runoff and streamflow forecasting in a snowmelt regime is important for snowmelt runoff principles, primarily through the use of hydrological modeling. Since this takes place in real-time, instead of involving careful analysis of historical data and repeated computer simulations, some aspects of snow hydrology must be treated differently than they are in design applications. In this paragraph, those facets of operational forecasting that pertain to snow hydrology will be discussed.

b. Short-term forecasting. For this discussion, short-term forecasting is defined as making streamflow predictions for several days into the future using observed and forecasted precipitation and temperature. In addition to generating a streamflow time series for a given basin, the forecast may also include a river-reservoir system simulation that produces an outlook of lake and reservoir elevations, river elevations, etc., all based upon the watershed-simulation input. The following summarizes some key points to be aware of in a snow environment.

(1) Model formulation. The possibilities for alternative model configurations have been discussed in Chapter 4. Although a relatively thorough and complex model is always to be considered, practical problems with the forecasting environment may dictate the use of a simpler formula than the one that may have been used for design analysis. Since snowmelt applications deal with considerable topographical relief, some ability to define the vertical distribution is highly desired. Situations where a vertically lumped model might be used are as follows.

(a) Rain-on-snow basins with relatively low-snow contribution.

(b) Basins that are relatively flat.

(c) Spring snowmelt basins where rain is a minor factor.

(2) Time increment. The computational time step will typically be defined by the basin size and is often 3 to 6 hr for rain-on-snow settings and somewhat longer for large spring runoff basins. For large basins, the interval should not exceed 12 hr for near-term forecasts, if the diurnal melt variation is to be described adequately.

(3) Snowmelt method. A temperature index is used almost exclusively for forecasting, although wind and other data can help guide the use of this index, as has been discussed in Chapter 6.

(4) Temperature input. For spring snowmelt simulations, temperature becomes the key variable defining melt quantities. A period-average temperature is usually used for forecast model input. In rain-on-snow settings, temperature is extremely important in establishing the freezing level, which in turn defines at what elevation precipitation will be falling as rain or snow. Temperature observations and forecasts will also be used to compute snowmelt for the forecast. Temperatures established for a station need to be projected to other elevations within the basin using a lapse rate that also is subject to change over time.

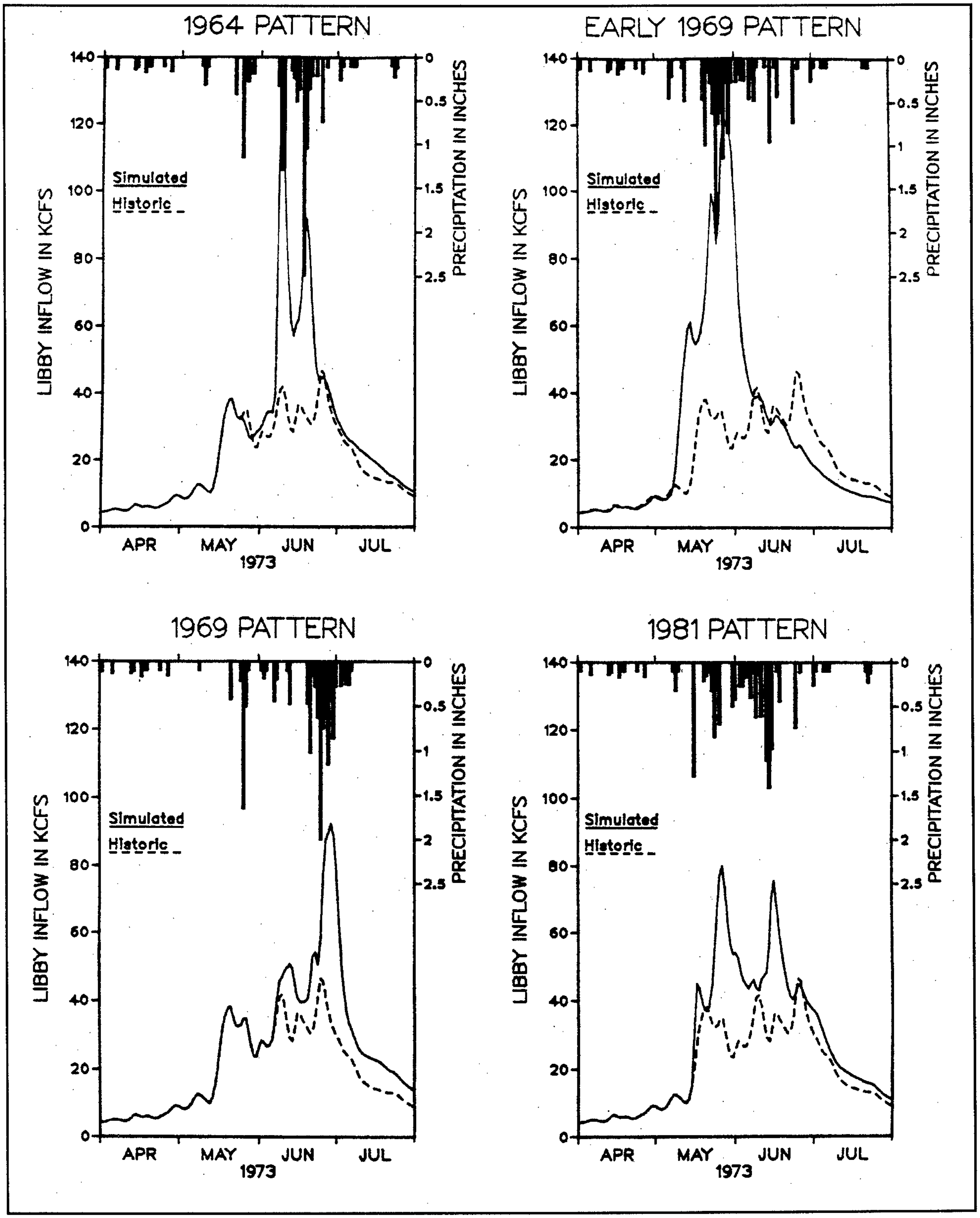

Figure 10-8. Hypothetical flood derivations, spring rain on snow

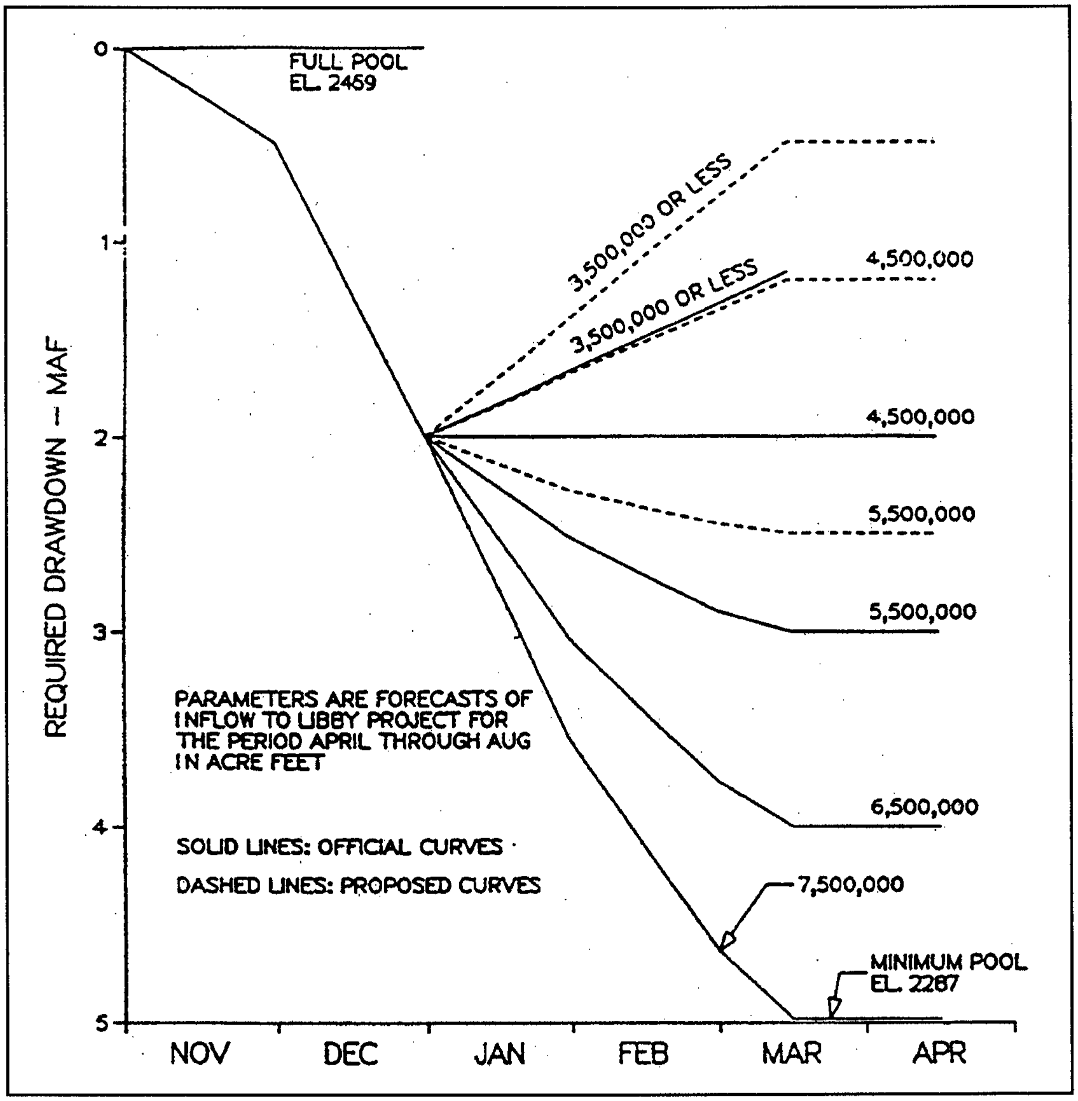

Figure 10-9. Comparison of original and revised rule curves, Libby project

(5) Rain input. For spring-summer flood basins, it may be possible to ignore light rainfall over snow-free areas, as discussed in Chapter 4.

(6) Snow-condition effects. This is often estimated intuitively by forecasters in a rain-on-snow setting rather than having it computed explicitly in the model. The effects on runoff are relatively small compared with other uncertainties, and they often occur early enough in the storm sequence so that they are of relatively minor importance for reservoir operations. In spring snowmelt, forecasting snow-condition effects are generally not a consideration.

c. Long-term forecasting. For this discussion, long-term forecasting is meant to include all forecasting extending beyond the above "short-term" definition. This would include seasonal-runoff-volume forecasts as well as streamflow forecasts extended over a long period of time. Since meteorological forecasting is not possible beyond several days into the future, long-term streamflow

forecasts need to reflect hypothetical or probabilistic input. A special case of this type of forecast is ESP forecasting, discussed separately below.

(1) Figure 10-10 portrays a forecasting procedure employed in the Columbia basin, wherein a long-term extension is applied to a short-term forecast. The input for the long-term forecast is a hypothetical temperature sequence that has been determined by analysis of historical meteorological data. Alternative sequences with different characteristics can be used. The extended forecast is useful in guiding the operation of large storage reservoirs that fill over the April-July snowmelt period:

(2) The following are additional guidance for long-term forecasting

(a) Model formulation. Since simulation over a long-term period is involved, a model capable of handling evapotranspiration and other long-term effects is required.

(b) Time increment. Because of the hypothetical nature of the results, a longer computation time step is sometimes employed during the extended period.

(c) Snowmelt method. Since the long-range forecast extends into the late summer, the snowmelt methodology must have provision for automatically changing melt-rate coefficients as the season progresses.

(d) Temperature input. This is provided as a hypothetical time series as shown in the above example or as a series of historical traces as used in the ESP technique (described below). The hypothetical series could represent subjectively derived patterns, historical temperature (and precipitation) from notable historic events, or a series developed by a relatively sophisticated stochastic analysis.

(e) Rain input. In the Columbia example, long-term rainfall is ignored because it is usually unimportant. The results are used with the understanding that they contain some volumetric bias because of this assumption.

d. Extended streamflow-prediction technique. This technique, developed and called ESP by the National Weather Service Office of Hydrology, is widely used by forecasting and management agencies throughout the United States. It is particularly applicable in a snowmelt regime where the long-term storage effect of accumulated snow results in a definite association with runoff several months later. It entails simulating a sampling of historical meteorological time series every time the forecast is made—20 or 30 years of data would typically be used. Producing a seasonal snowmelt runoff forecast is illustrated in Figures 10-11 and 10-12. Early in the snow accumulation season, relatively little information about the year being forecasted has yet to be known, since only a small portion of the precipitation has accumulated for the year. The resulting display of model results has a large variance, not unlike the historical sampling of runoff data itself. As the season progresses, later forecasts take on the specifics of the year in question, and future variance created by the range of future meteorological possibilities diminishes. The ESP technique offers several advantages over other techniques in long-range forecasting.

- It is relatively rigorous, statistically.

- It permits a wide range of forecast products, including volume and peak flows.

- It provides error statistics and displays.

The drawback of the technique is that it uses considerable computer resources. On a large river with many subbasins, this drawback may preclude its use. ESP procedures require a continuous soil moisture accounting model that can operate through snow-accumulation periods as well as through extended periods of snowmelt.

10-5. Snow Modeling Considerations in Continuous Simulation

a. Overview. Continuous soil-moisture-accounting modeling is used regularly in snowmelt regimes, particularly in ESP forecasting and operational analysis. Because this modeling extends over long times, including the snow-accumulation period, additional facets of snow hydrology need to be considered beyond what is dealt with when modeling snowmelt only. The simulation process during snow

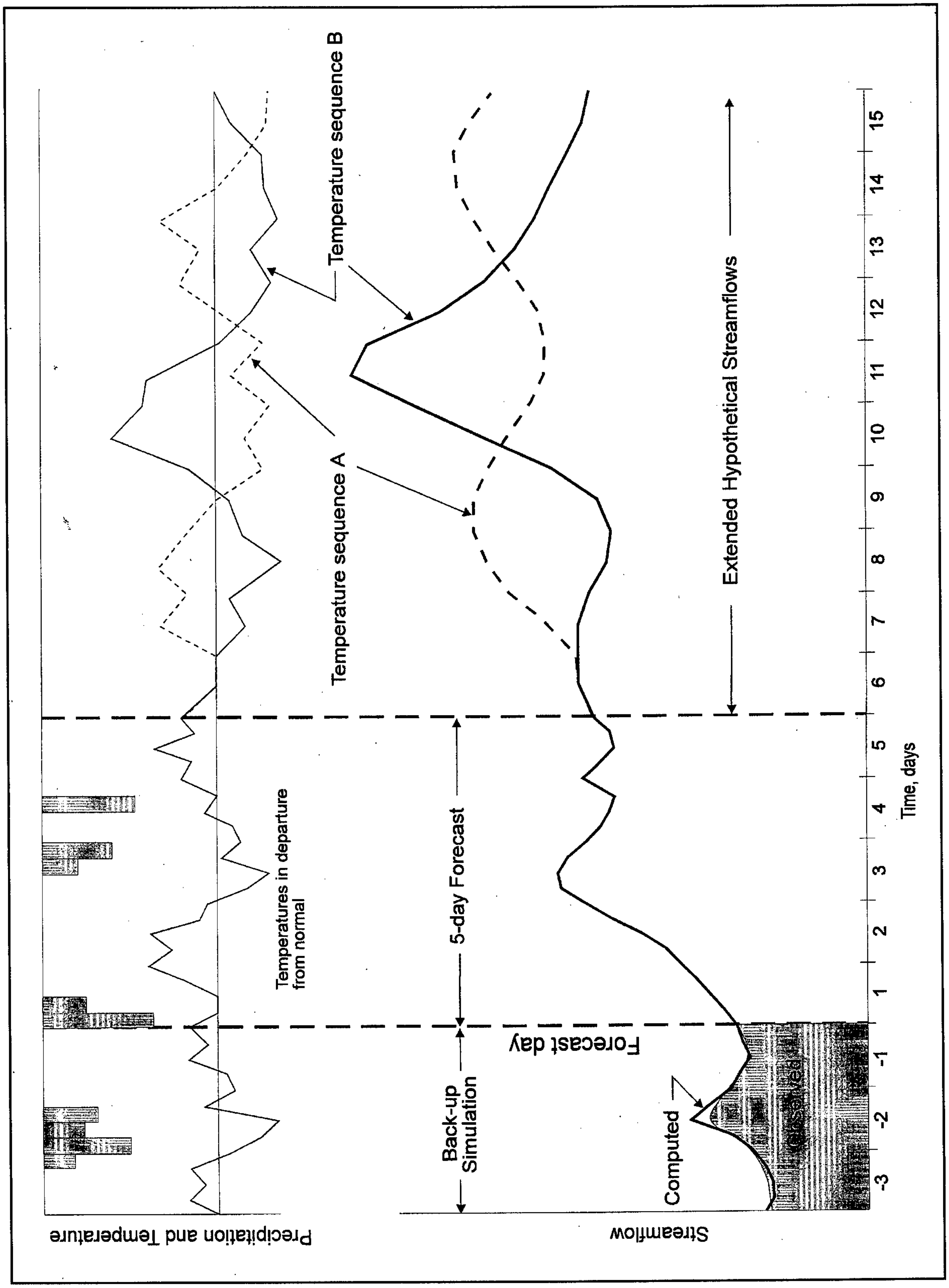

Figure 10-10. Example of short- and long-range streamflow forecasts

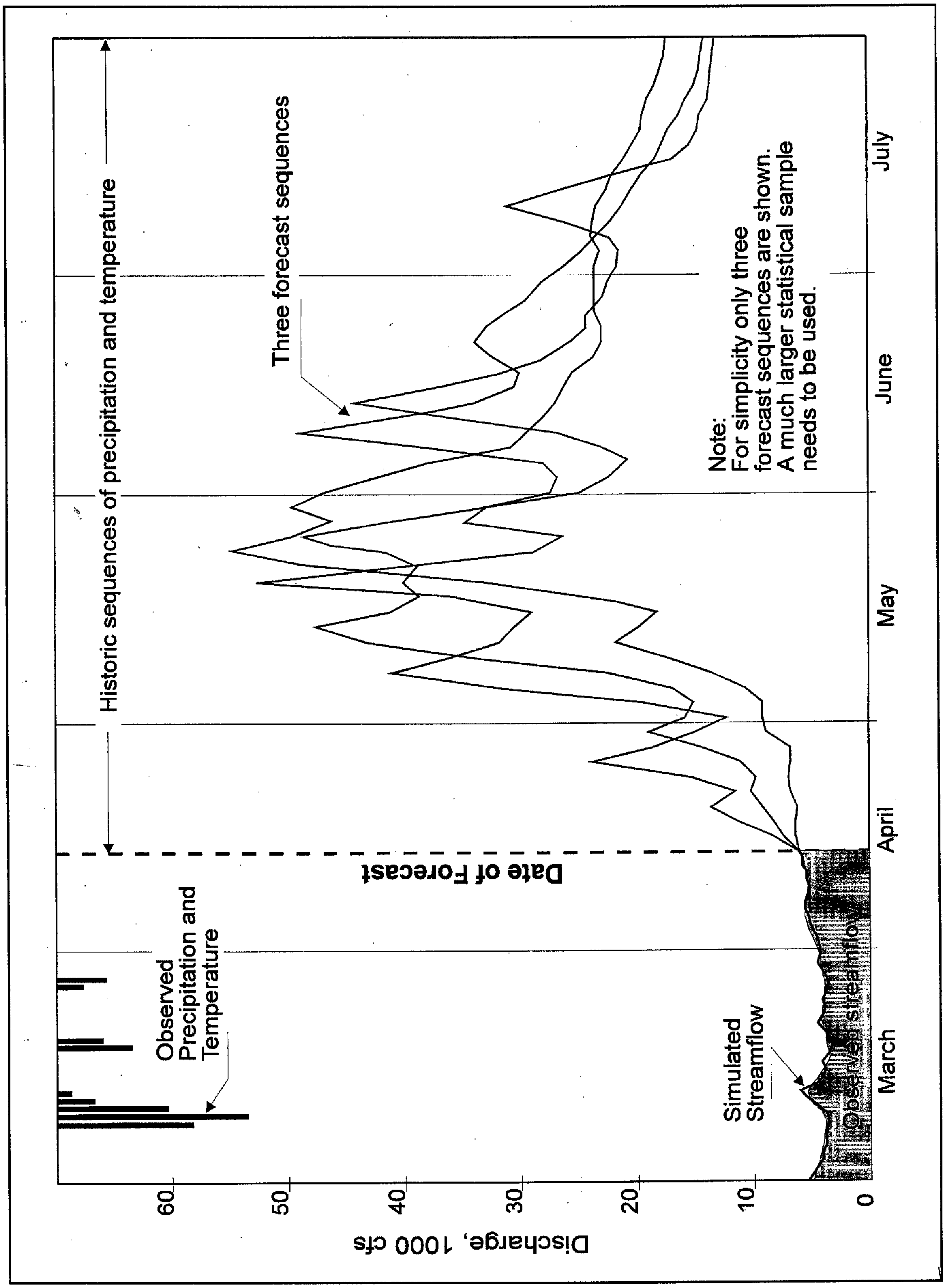

Figure 10-11. Hydrographs generated with the ESP technique (continued)

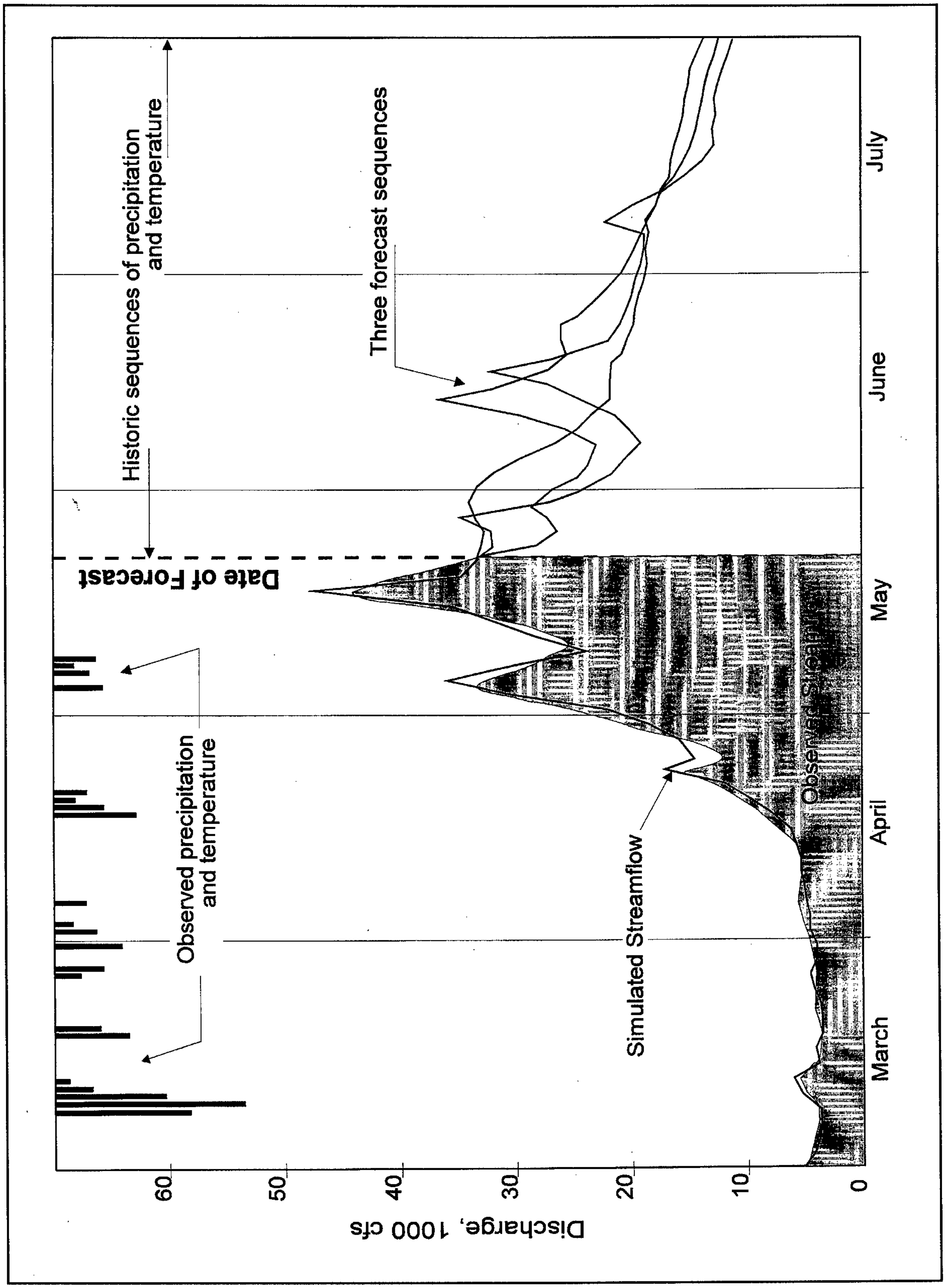

Figure 10-11. (concluded)

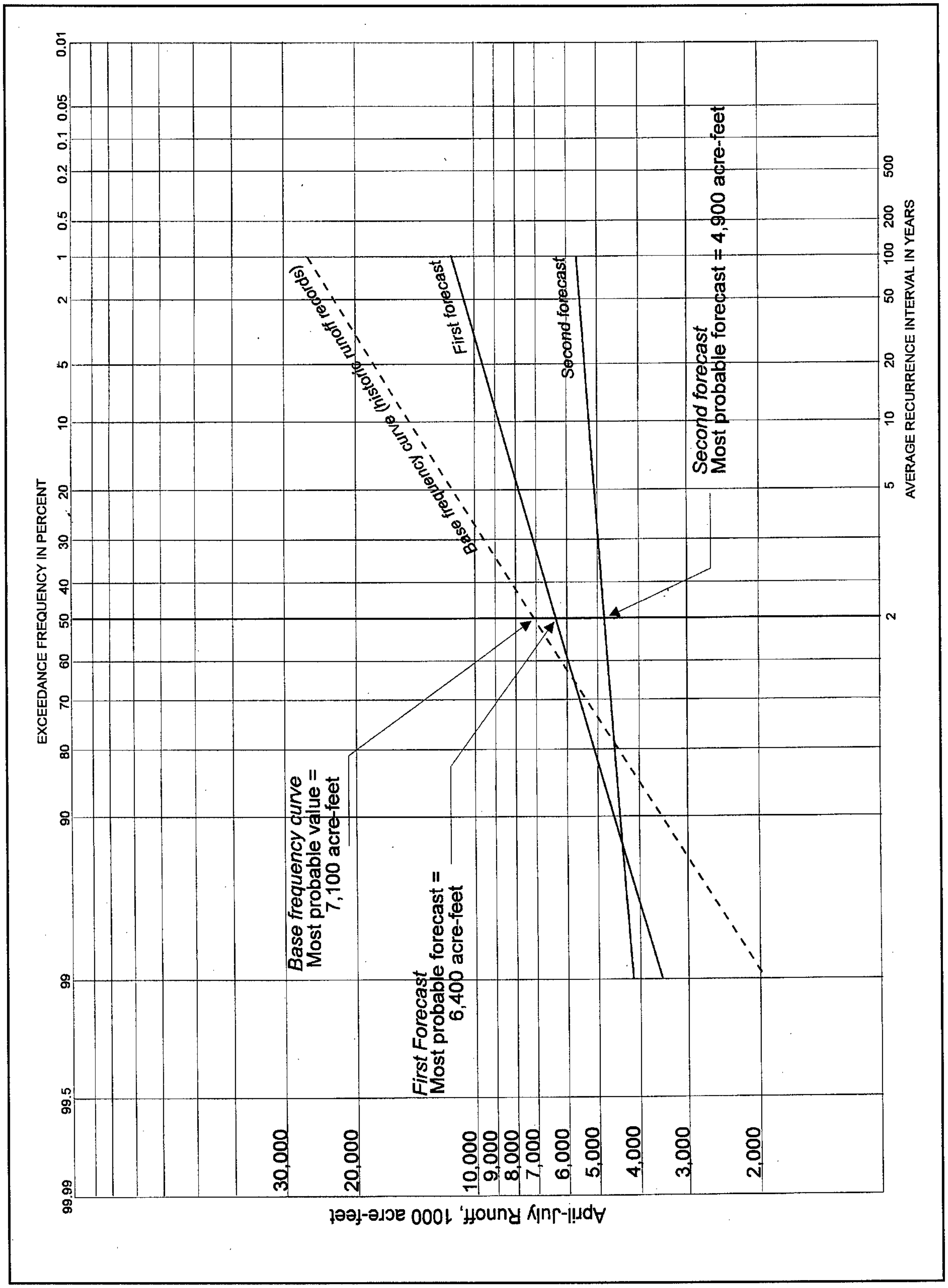

Figure 10-12. Statistical analysis associated with the ESP technique

accumulation and melt is illustrated by the algorithms shown in Figures 4-4 and 10-13, assuming an elevation-band model. Calibration of a continuous model typically uses a continuous period of data for many years, if not the entire period of record. The calibration must consider the long-term volumetric effects and seasonal water balance, along with the general ability to reproduce streamflow without bias. For snowmelt environments, the input variables are precipitation and air temperature (station maximum and minimums for a daily time step). The winter-snow accumulation is computed by the model. Observed snow measurements could be used as an additional means for judging the model calibration if desired.

b. Simulation guidance. The following summarizes some factors that need to be considered with this method of modeling.

(1) Time increment. Since the model operates through flood as well as low-flow periods, some models provide for an automatically changing computational period based upon rate of change of input.

(2) Snowmelt method. The temperature-index approach is essentially a requirement since such a large amount of historical data are employed. The model must be able to compute melt-rate coefficients as a seasonal variable. Melt from ground conduction could be added as melt source because of the extended computational periods involved.

(3) Temperature input. Temperature data are exclusively historical station data, generally input as daily maximums and minimums. These must be converted to area mean values through some form of weighting process, and it is desirable to have flexibility to vary the temperature weighting seasonally. A temperature station may, for instance, index an area's temperature differently during winter storms than it does during summer melt under clear skies. Air temperatures must also be lapsed to the appropriate elevation. A fixed lapse rate is typically used, although this could be made to vary seasonally also.

(4) Rain input. Historical station data are used as input, so a conversion to area means is required. As with air-temperature data, the conversion process should have some flexibility to consider seasonal variations. A factor to consider is that different gauge catch efficiencies result when precipitation is snow versus rain.

(5) Interception, evapotranspiration, and sublimation. These factors must be simulated, using whatever algorithm is available in the model. Temperature is usually the independent variable used to compute evapotranspiration. Sublimation of snow must also be accounted for, since this can be a significant loss over extended periods of time.

(6) Snow-condition effects. Continuous simulation modeling needs to account for these phenomena explicitly. A sample algorithm for this process has been presented in Chapter 7.

(7) Glacial melt. For areas having continental glaciers, melt from this source can be significant in late summer. If a specific glacier-melt routine is not provided in a model, this phenomenon could be represented by treating the glacial areas as separate subbasins and creating special characteristics using a standard model.

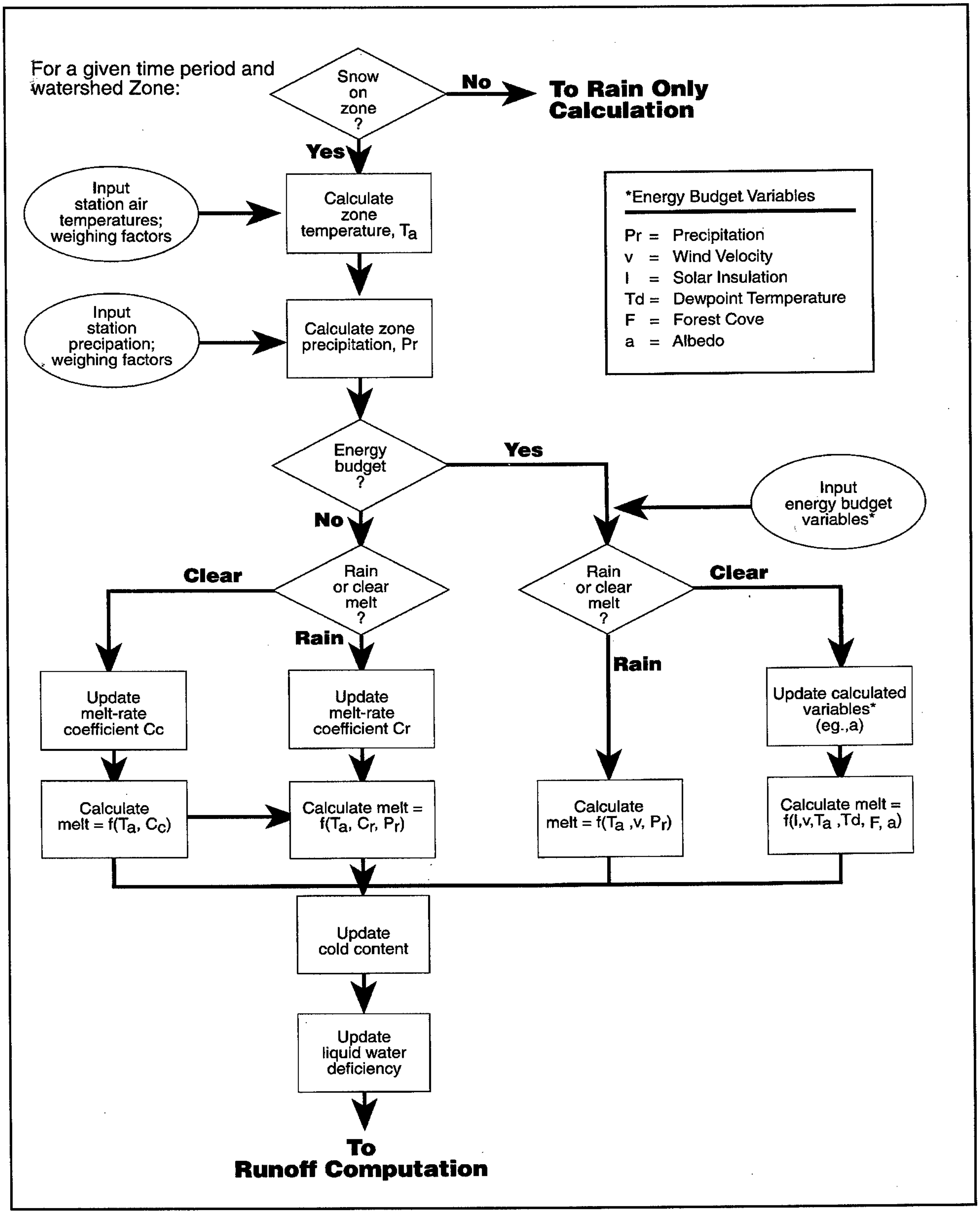

Figure 10-13. Algorithm of snowmelt simulation, continuous simulation model

Chapter 11
Guidelines for Snowmelt Model Selection

11-1. General Introduction

The aim of the hydrologist in the choice of a particular mathematical modeling scheme depends on a clear definition of the problem to be solved and upon the database that is available to describe the physical system (Anderson and Burt 1985). The preceding chapters have discussed the fundamental physical and engineering processes that need to be addressed and the database requirements in snowmelt modeling. The key points in the selection of the appropriate modeling methodology are as follows.

- Operation and calibration data availability.

- Expected physiographic and climatic conditions.

- Detail and type of results required.

- Probability of extreme events.

a. The availability of operation and calibration data is a key constraint to the choice of methodology. If an ungauged catchment is the area of interest, any model involving optimization procedures based on historical discharge record or a complex conceptual energy budget would be ruled out because of the absence of data. The accuracy, representativeness, and validity of the collected data are as important as their availability in model selection. Models based on physical parameters require physically meaningful data inputs to correctly characterize the snowmelt process. Even with simple empirically derived index methods, the issues related to data reliability are of major importance. The versatility of a model in characterizing varying physiographic and climatic conditions is an important factor. This is called model mobility and is critical to applying a model to a new site. Most calibrated snowmelt models tend to be site-specific, and their applicability to differing conditions is a function of their deterministic quality. The purpose of the analysis is probably the most exact requirement of snowmelt analysis. Whether or not the model is used for real-time forecasting is also a consideration. The detail and type of results required, e.g., peak flow, event volume, event hydrograph, or a long-term sequence of flows, weigh greatly on the choice of the appropriate modeling scheme.

b. The probability of extreme events leads the hydrologist to consider a physically based approach versus empirically derived indexes. As mentioned previously, index methods are most accurate under normal conditions, whereas energy budget approaches, owing to their physical basis, are more accurate at forecasting extreme events.

c. For the operational hydrologist, the availability of resources and time to carry out a snowmelt-forecasting analysis is of extreme importance. Some techniques, such as a complete energy budget approach to snowmelt analysis, require extensive commitments of personnel, computer resources, and expertise to become operational. These management applications or operational constraints need to be fully considered in selecting methodology. In general, two main issues emerge in model selection: the need for widely applicable models and the requirement for suitable databases to support the snowmelt modeling.

11-2. Specifics of Snowmelt Model Selection

As mentioned previously in Chapter 10 (the analysis alternatives are summarized in Table 10-1), numerous alternatives are available for approaching computing snowmelt in hydrological engineering analysis and forecasting. Table 11-1 lists the characteristics of six operational snowmelt models that have been chosen because they are applied by USACE, generally in North America. These models are used by Federal, State, and private institutions. The USACE hydrologist should be aware of the framework of other agencies' models as they pertain to operation of USACE projects.

a. The USACE models, SSARR and HEC-1, are typically used for snowmelt. The choice between the two models, for example, might be based on the need for short- or long-term forecasts. The SSARR model is a continuous simulation model that does continuous accounting of snowpack conditions, whereas HEC-1 is an event-based model that does not have snowmelt accounting. Therefore, if the engineering applications

Table 11-1
Comparison of Operational Snowmelt Models (After Schroeter 1988; Ontario Ministry of Natural Resources 1989)

	Model Name / Type					
	SSARR C*	HEC-1 E*	NWSRFS C	PRMS C	SRM E	GAWSER E
Energy budget	o	o	•	•		•
			(rain on snow)			
Modeled components						
Temp. index	•	•	•		•	
Elev. correction	•	•	o	•	•	
Areal snow cover	•	•	•	•	•	•
Forest/open	o	o		•		•
Heat deficit	•		•	•		•
Water storage	•		•	•		•
Density depth		o				•
Frozen ground		o	o			
Input data requirements						
P	•	•	•	•	•	•
T_a	•	•	•	•	•	•
T_d	o	o				•
u_z	o	o				•
Q_{sin}	o			•		•

Note: • = standard; o = optional; C = continuous-simulation capacity; E = single-event model; P = precipitation; T_a= air temperature; T_d= dew point; u_z = wind speed; and Q_{sin} = incoming solar radiation.

require a short-term forecast, the hydrologist might choose HEC-1, and for long-term forecasts, SSARR.

b. The other models listed are for other agencies and institutions. The National Weather Service, as the primary U.S. river forecast agency, uses the National Weather Service River Forecast System (NWSRFS), which is an offspring of the Stanford Watershed Model (Anderson 1973). PRMS is supported by the U.S. Geological Survey and employs new technologies for distributing runoff based on hydrological response units (Leavesley et al. 1983). The Agricultural Research Service (Martinec, Rango, and Major 1983) supports the model SRM. It has been applied worldwide and consists of a simple, rational-form-based runoff model. The use of satellites to remotely sense snow-covered area to derive snow cover depletion curves is an important feature of this model. The last model listed in Table 11-1 is Guelph All-Weather Storm-Event Runoff (GAWSER) (Schroeter 1989). It is a Canadian model that has been applied operationally. The features that might affect its applicability are its distributed nature and its use in prairie, agricultural regions. In the following (Paragraph 11-3), summary fact sheets for each model are provided for quick reference to the models, and in Appendix F, a more complete description of each model is detailed. By using Table 11-1 and these fact sheets, the general capabilities of these models can be seen, and an appropriate snowmelt model can be selected.

11-3. Summary Fact Sheets for Selected Snowmelt Models

a. Model name, Streamflow Synthesis and Reservoir Regulation Model (SSARR).

(1) Description. Continuous streamflow simulation model using either a lumped parameter or distributed (elevation band) representation. SSARR contains a watershed model and a river system and reservoir regulation model. Originally developed in 1956, it has been successfully implemented for numerous diverse river basins worldwide. Model routing in the watershed and river system is accomplished by cascading linear reservoirs. Evapotranspiration is computed as a function of air temperature or from input-evaporation data. The model has been used for both short-term and long-term forecasting, including ESP-type forecasts.

(2) Snowmelt routine description. Two options:

(a) Temperature-index method with lapse-rate correction.

(b) Generalized energy budget snowmelt equation (USACE 1956). Daily melt is calculated and distributed throughout the day using distributions based on the diurnal fluctuations of heat supply for melting snow. Areal distribution of snow is by means of a snow cover depletion function or by elevation bands. Ground melt is available.

(3) Suitability and restrictions. Suitable to a wide range of basins; flexible in time step and basin size. Does not deal directly with occurrence of frozen ground; limited successful application to permafrost conditions. Lumped snowmelt relationships only allow for elevation-affected snow distribution and melt.

(4) Source.

U.S. Army Corps of Engineers
North Pacific Division, CENPDEN-WM
PO Box 2870
Portland, OR 97208

(5) Documentation. U.S. Army Corps of Engineers, User Manual, SSARR Model, Streamflow Synthesis and Reservoir, North Pacific Division, January 1991.

b. Model name, HEC-1, HEC-1f.

(1) Description. Event-based simulation model. Flexible component package to simulate surface runoff response to precipitation or snowmelt for complex, multisubbasin, and multichannel river basins. HEC-1f is a version used for real-time flood forecasting. Runoff transformation is done by unit hydrograph, with several options being available.

(2) Snowmelt routine description. Two options:

(a) Temperature-index method. Snow distribution specified by elevation bands.

(b) Energy budget snowmelt equation (USACE 1956) available for design analysis.

(3) Suitability and restrictions. Fully supported for use with HEC Data Storage System. Flexible in choice of watershed routing functions. Restricted by lack of soil and snow-moisture accounting routings. No accounting for frozen ground.

(4) Source.

Hydrologic Engineering Center
U.S. Army Corps of Engineers
609 Second Street
Davis, CA 95616

(5) Documentation. U.S. Army Corps of Engineers, HEC-1, Flood Hydrograph Package, User's Manual, Hydrologic Engineering Center, Davis, California, September 1990.

c. Model name, National Weather Service Snow Accumulation and Ablation System (NWSRFS)

(1) Description. Incorporating the Sacramento Watershed Model and other hydrology computation

modules, NWSRFS was developed in 1972 at the Hydrologic Research Laboratory of the NWS Office of Hydrology. It can continuously simulate watershed response for flood forecasting. Accounts for soil moisture among five reservoirs, differentiating between free and capillary water. Runoff transformation done by unit hydrograph.

(2) Snowmelt routine description. Snowmelt routine consists of two general sectors: a meltwater production unit and a meltwater storage and transmission component. During rainless periods, temperature index using a seasonally adjusted melt factor is used. During rain or snow events, a simplified energy budget approach is used, which requires only air temperature and precipitation data. Heat deficit of the snowpack is also continuously monitored.

(3) Suitability and restrictions. Has been applied to more than 20 basins in the United States over a wide range of climatic and snow cover conditions. Developers have designed and tested a snow energy budget model (Anderson 1979) and frozen ground routine (Anderson and Neuman 1984), which are being implemented.

(4) Source.

Office of Hydrology, W23
National Weather Service, NOAA
8060 13th Street
Silver Spring, MD 20910

(5) Documentation. Anderson, Eric A., National Weather Service River Forecast System—Snow Accumulation and Ablation Model, NOAA Technical Memorandum NWS 17, U.S. Dept. of Commerce, Silver Spring, Maryland, 1973.

d. Model name, Precipitation-Runoff Modeling System (PRMS).

(1) Description. Multipurpose model for short- and long-term forecasting of daily streamflow from snowmelt. Originally developed for mountainous areas, it has been recently and successfully applied throughout the U.S. Basin and is divided into HRUs. Used primarily for watershed analysis.

(2) Snowmelt routine description. Two-layered snowpack energy budget for each HRU (lumped processes within). Heat transfer by conduction within layers.

(3) Suitability and restrictions. Well suited for short-term forecasts (3 to 5 days) of mean daily discharge. Use of HRUs well founded in physical process modeling. No soil-moisture or frozen-ground accounting.

(4) Source.

U.S. Geological Survey
Water Resources Division
MS 412 Box 25046
Denver Federal Center
Denver, CO 80225

(5) Documentation. Leavesley, G. H., Lichty, R. W., Troutman, B.M. and Saindou, L. G., Precipitation-runoff Modeling System, User's Manual, U.S. Geological Survey Water Resources Investigators Report B3-4238, 1983.

e. Model name, Snowmelt Runoff Model (SRM).

(1) Description. First developed by Dr. J. Martinec, Federal Institute for Snow and Avalanche Research, Davos, Switzerland, and first used in 1973. Originally developed to make use of remotely sensed snow cover data, SRM has been applied to a wide range of basins.

(2) Snowmelt routine description. Snowmelt is calculated using the temperature-index method, employing precipitation, air temperature, and depletion curves of snow cover derived from ground-based data or Landsat. No accounting for snow properties and uses rational form for transforming snowmelt to discharge. Spatial distribution accounted for using elevation bands.

(3) Suitability and restrictions. Suitable for mountainous basins less than 4000 km^2. Limited to daily discharge calculations and no soil moisture accounting. Well suited for modeling when only data source is remotely sensed snow cover information.

(4) Source.

Dr. A. Rango
Hydrology Laboratory
Agricultural Research Service
Building 007, Rm. 139
Beltsville, MD 20705

(5) Documentation. Martinec, J., Rango, A., and Major, E. The Snowmelt Runoff Model (SRM) User's Manual, NASA Reference Publication 1100, Washington, DC, 1983.

f. Model name, Guelph All-Weather Storm-Event Runoff Model (GAWSER).

(1) Description. Modified version of HYMO and is a deterministic event-based model. Originally designed for agricultural areas, has been recently interfaced to a distributed snow model (Areal Snow Accumulation-Ablation Model, see description). Has options that deal with distributed soil characteristics. Has been used for operational forecasting in Canada.

(2) Snowmelt routine description. Temperature-index approach to determine snowmelt. Refreeze, compaction, new snow deposition, rain deposition, snowmelt, and release of liquid water are considered. Recently added cell-based detailed energy balance to account for areal variability of snow cover within subwatershed.

(3) Suitability and restrictions. Model originally designed for agricultural areas and has data requirements that restrict usefulness to areas with high data availability.

(4) Source.

School of Engineering
University of Guelph
Schroeter and Associates
Grand River Conservation Authority

(5) Documentation: Schroeter, H., GAWSER Training Guide and Reference Manual, Grand River Conservation Authority (GRCA), October 1989.

Appendix A
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Appendix B
Glossary and Notation

B-1. Glossary

The following is a summary of terms used in this manual that are common to snow hydrology and its related fields.

Albedo
The ratio of the amount of shortwave radiation reflected by a surface to the total flux incident to the surface.

Back-radiation
Long-wave (terrestrial) radiation emanating from clouds, forest canopy, atmospheric particles, etc., and directed towards the earth.

Blackbody (radiation)
A body that radiates for every wavelength the maximum intensity of radiation possible for a given temperature. (The term does not imply that the radiating substance is colored black.)

Calorie (gram-calorie)
The amount of heat required to raise a gram of water 1 °C, from 14.5 to 15.5 °C.

Cold content
The amount of energy required to raise a snowpack to 0 °C, expressed in terms of the amount of water needed to be produced at the surface to release energy by freezing. Applied primarily to determine initial losses during wintertime rain on snow.

Condensation
Heat energy (and snowmelt) produced through the phase change of water from a vapor to a liquid.

Conduction
Heat energy (and snowmelt) produced by heat transferred through a solid body by molecular activity. Applied to heat conducted from the ground in snow hydrology.

Continuous simulation
Simulation with a generalized hydrological model in which the model is operated continuously through dry as well as storm periods. Requires the ability to simulate evapotranspiration as well as other phenomena.

Convection
Heat energy (and snowmelt) produced by the transfer of heat through the movement of the air (or any fluid), brought about by natural or induced pressure differences. Also called sensible heat transfer.

Degree-day factor
See *melt-rate coefficient.*

Dew point
The temperature to which the air must be cooled—at constant pressure without the removal or addition of moisture—to produce condensation of water vapor.

Distributed (parameter) model
A category of conceptual models, in which the watershed parameters are defined by breaking down the total basin into smaller, independently computed subunits. This leads to an improved and more physically based model definition as compared with a *Lumped model.*

Elevation bands
Zones of equal elevation in a watershed model. One method of achieving a degree of distribution in defining a hydrological model of a basin.

Energy budget
A method of snowmelt analysis and simulation for which the energy flux components are explicitly accounted.

Energy flux
The rate of change of energy (e.g., shortwave radiation) per unit time.

Extended streamflow prediction (ESP)
A forecasting technique in which future conditions are simulated by using for future input (e.g., precipitation and temperature) a number of historical time series, all beginning with the model's current state (e.g., soil moisture, snowpack).

Gravitational water
Liquid water in a snowpack that is in transit through the pack under the influence of gravity.

Ground conduction melt
See *Conduction.*

Hygroscopic water
Liquid water held in the snowpack crystal matrix that is not available for runoff until the snow crystals have melted.

Hypothetical floods
Simulated floods used for design, in which the magnitude is typically expressed in terms of probability of occurrence or as a maximum probable event.

Incident radiation
Solar radiation that falls on a surface.

Insolation
Total solar radiation flux received on a horizontal surface.

Joule
A measure of heat energy or work in the SI system of units, equal to one watt per second. One gram-calorie equals 4.186 joules.

Langley
A measure of solar radiation equal to one calorie per square centimeter.

Latent heat
The heat quantity taken in or given off when a substance changes its state, e.g., from liquid to gas.

Liquid-water holding capacity
The capacity of a snowpack to retain nongravitational liquid water.

Long-wave radiation
Radiation energy from terrestrial sources, occurring with wavelength of 6.8 to 100 µm. Also called *Thermal radiation.*

Lumped model
A conceptual model in which a single set of parameters defines the system. See also *Distributed model.*

Melt-Rate coefficient
A coefficient used in the *Temperature index* equation for snowmelt. Also called a *Degree-day factor.*

Metamorphism
The change in the character of a snowpack as it matures, in which individual crystals become rounded and bound together and the snowpack becomes more dense and is warmed to 0 °C.

Precipitation
Rain, snow, hail, etc., falling to the ground.

Primed snowpack
A mature snowpack in which the temperature of the snow has become isothermal at 0 °C and the liquid water deficiency is satisfied, and is ready for runoff-producing melt.

Radioisotopic gauge
A method of measuring *Snow water equivalent* (SWE) by sensing the attenuation of radiation emitted from a source.

Rain melt
Snowmelt produced by the heat given up after rainwater has fallen on the snowpack.

Relative humidity
The ratio of the water vapor content of the air compared with the saturated content at the same temperature. It can be computed by dividing the actual vapor pressure by the saturated vapor pressure.

Residual
In correlation analysis, the difference between the predicted and observed value of the independent variable.

Ripeness
The degree of maturity of a snowpack as measured by the internal temperature, character of the snow crystals, and liquid-water content.

Saturated vapor pressure
In meteorology, the vapor pressure when the air has reached its capacity for water vapor; it is saturated. This is a function of air temperature. See *Vapor pressure*.

Sensible heat melt
See *Convection*.

Shortwave radiation
Radiation emitted by the sun, with wavelength of 0.2 to 2.2 μm.

SNOTEL
Acronym for SNow TELemetry system, an automated snow data collection system managed by the U.S. Natural Resource Service in the western United States.

Snow
The form of precipitation that falls as ice in a crystalline form, each crystal having a unique shape, with sharply defined edges and abrupt points.

Snow condition
A relative measure of a snowpack's degree of *Metamorphism*, as it changes from a fresh, dry state to a mature, *Ripe* state. Applied to *Cold content* determinations.

Snow cover depletion curve
A curve that defines the percentage of areal snow cover of a basin as a function of percent of total anticipated runoff. Used for estimating snow cover and snowline elevation in simulation models.

Snow course
A manual snow-sampling station at which several *Snow tube* samples are taken to get representative values of depth, density, and SWE.

Snow density
Theoretically, the mass of a unit volume of snow, expressed in kilograms/cubic meter. More commonly, it is expressed as a percentage—for a unit area, the depth of the *SWE* divided by the depth of the snow (10% = 100 kg/m^3).

Snow pillow
A device that automatically measures the snowpack *SWE*, consisting of a rubber or stainless steel pillow filled with liquid.

Snow survey
A general term for the manual sampling of snow.

Snow water equivalent (SWE)
The liquid-water equivalent of the snowpack, expressed in terms of depth.

Snowfall
The depth of newly fallen snow, measured before it becomes compacted.

Solar constant
The radiant solar energy flux received outside the Earth's atmosphere, on a surface normal to the sun's rays. Established at 1.365 kW/m^2.

Solar radiation
Radiation emitted by the sun. See *Shortwave radiation*.

Stefan-Boltzmann equation
A fundamental relationship that states that energy radiated by a blackbody is equal to the fourth power of its Kelvin temperature times the Stefan-Boltzmann constant.

Temperature index
A simplified method of computing snowmelt in which air temperature is used to index all the energy sources involved.

Terrestrial radiation
See *Long-wave radiation*.

Thermal quality
Ratio of heat required to melt a unit mass of snow to that of ice at 0 °C.

Thermal radiation
See *Long-wave radiation.*

Turbulent transfer/exchange
The physical mechanism occurring in the 2 to 3 m (6 to 10 ft) of the atmosphere immediately above the snow surface by which sensible and latent heat energy fluxes are transferred to the snow surface.

Vapor pressure
In meteorological applications, the partial pressure exerted by water vapor in the atmosphere, expressed in millibars or millimeters of mercury. This is an absolute measure of the amount of water vapor in the air. See *Saturated vapor pressure.*

B-2. Notation

The following is a listing of notations used in the equations presented in this manual. Widely known and accepted notations (e.g., meters, kilograms) are not included. Since both SI and English units are used in this manual, both systems could be shown for most variables; however, where one convention has been used exclusively in the manual, only those units are shown.

ϵ emissivity of snow, decimal fraction

ρ_s density of snow, g/cc, kg/m^3, percent

ρ_w density of water, kg/m^3

σ Stefan-Boltzmann constant, kJ/m^2 s K^4 or ly/min K^4

A cross section area

a snow surface albedo, decimal fraction

B thermal quality of snow, decimal fraction

BF base-flow index in runoff-volume forecast equation, units of depth

cal calorie

CF_a correction factor for temperature-measurement height adjustment, decimal fraction

CF_b correction factor for wind-velocity measurement height adjustment, decimal fraction

C_m melt-rate coefficient in temperature-index equation, inches/degree-day

C_p specific heat of water, kJ/kg °C

C_r conversion factor, cold-content simulation equation, inches/degree-day

d depth of snow, inches or centimeters

df degrees of freedom

e_a vapor pressure of air, millibars

e_s saturation vapor pressure of air, millibars

F basin forest-canopy cover shading from shortwave radiation, decimal fraction

FP fall index in runoff-volume forecast, units of depth

I_i solar insolation flux, ly/day, mJ/day m^2, W/m^2

J joule

K Kelvins

k basin wind exposure factor in energy budget equation, decimal fraction

k' basin shortwave radiation melt factor in energy budget equation, decimal fraction

kJ kilo-joules

L latent heat, kJ/kg or cal/g

ly langley

M combined melt from all energy sources, inches or millimeters

M_c snowmelt due to latent heat of condensation, inches or millimeters

M_{ce} combined snowmelt, condensation, and convection, inches or millimeters

M_e snowmelt due to convection heat transfer, inches or millimeters

M_l snowmelt due to long-wave radiation heat, inches or millimeters

M_r snowmelt due to heat released from rainwater, inches or millimeters

M_s snowmelt due to shortwave radiation heat, inches or millimeters

N cloud cover, decimal fraction

p atmospheric pressure at location

p_0 atmospheric pressure at sea level

P_r daily rainfall, inches or millimeters

Q heat energy (general), typically kJ/m^2·day, mJ/m day, ly/day, W/m^2

Q_c energy flux from condensation

Q_e energy flux from convection from the air

Q_g energy flux from ground conduction

Q_i internal energy in snowpack

Q_{lb} long-wave back (towards the earth) radiation flux

Q_l net long-wave radiation

Q_m total heat energy flux available to produce snowmelt

Q_r energy flux from rainwater

Q_s energy flux from shortwave radiation

S SWE index in runoff-volume forecast equation

SP spring precipitation index in runoff-volume forecast equation

SWE snow water equivalent, inches or mm

T_a air temperature, °C or °F

T_b base temperature in temperature-index equation, °C or °F

T'_c difference between cloud and snow surface temperatures, °C or °F

T'_d difference between dew point and snow surface temperatures, °C or °F

T'_s snow-temperature deficit below freezing, °C or °F

T_d dew-point temperature, °C or °F

T_r temperature of rain, °C or °F

T_s temperature of snow, °C or °F

v wind velocity, mph or km/hour

W watt

W_c cold content, inches

WP winter-precipitation index in runoff-volume forecast equation

Y seasonal runoff volume (dependent variable) in runoff-volume forecast equation

z_a height of temperature measurement, feet or meters

z_b height of wind velocity measurement, feet or meters

Appendix C
Summary of Basic Physics Principles—Heat, Heat Transfer, and Thermal Properties of Water

C-1. Temperature

Table C-1 compares the temperature scales for the three conventions used for snow hydrology fundamentals.

Table C-1
Comparison of Temperature Scales

	Celsius °C	Fahrenheit °F	Kelvin K
Melting point of ice	0	32	273
Boiling point of water	100	212	373
Divisions between fixed points	100	180	100

Conversion formulas: $$°F = \frac{9}{5}°C + 32$$

$$°C = \frac{5}{9}(°F - 32)$$

$$K = °C + 273$$

C-2. Heat Energy

In older literature, heat quantity was expressed in terms of calories, where

one g-cal = heat required to raise 1 g of water 1 °C, from 15 to 16 °C

In expressing mechanical energy, the convention in the metric system is to use Joules or ergs (1 erg = 10^{7}J). Recognizing that heat is a form of energy, the calorie is now defined in terms of the joule. A joule is a unit of work energy equal to a newton-meter. A watt, a unit of power, is equal to one joule per second. By international agreement

$$1 \text{ g-cal} = 4.186 \text{ J}$$

In hydrology and meteorological practice, the term kilo-Joule is used

$$1 \text{ kg-cal} = 4.186 \text{ kJ}$$

Table C-2 summarizes the equivalents of energy/work for several contemporary and older standards of units.

Table C-2
Units of Energy and Work

	J	kcal	kWh	Btu	ft-lb
1 J =	1	239×10^{-6}	277.8×10^{-9}	948.4×10^{-6}	0.7376
1 kcal =	4186	1	1.163×10^{-3}	3.968	3.087×10^{3}
1 kWh =	3.6×10^{6}	860	1	3413	2.655×10^{6}
1 Btu =	1055	0.252	293×10^{-6}	1	778.6
1 ft-lb =	1.356	324×10^{-6}	376.8×10^{-9}	1.286×10^{-3}	1

Note: J = Joule (1 Joule = 1 watt-second); kcal = 1000 calories; kWh = kilowatt-hour; Btu = British Thermal Unit; ft-lb = foot-pound.

C-3. Heat Capacity, Specific Heat

The ratio of heat supplied a material to the corresponding temperature rise is called the heat capacity.

$$\text{Heat Capacity} = \frac{Q}{\Delta t}$$

To obtain a figure that is characteristic of the material of which the body is composed, the specific heat of a material is used. This is defined as heat capacity per unit mass

$$C_p = \frac{\text{heat capacity}}{\text{mass}} = \frac{Q}{m\Delta t}$$

where C_p equals the specific heat, commonly expressed in kJ/(kg K) or cal/(g °C).

Table C-3 lists specific heats for substances common to snow hydrology.

Table C-3
Common Specific Heats

Substance	kJ/(kg K)	cal/(g·°C)
Water, 0 °C	4.217	1.01
Water, 20 °C	4.182	1.00
Ice	2.09	0.55
Air, dry, 20 °C	1.007	0.24
Sat. water vapor, 0 °C	1.864	0.46

C-4. Change in Phase, Latent Heat

The amount of heat absorbed (or given off) by a mass of material as it undergoes a change in phase (solid to liquid to gas, or reverse) is called the latent heat. The phase change involved occurs without a change in the temperature of the material itself. To compute the heat requirement for a phase change:

$$Q = mL$$

where

Q = heat energy, kJ (or cal)

m = mass, kg (g)

L = latent heat, kJ/kg (or cal/g)

a. Water and other substances can undergo a direct phase change from solid to gas when conditions are favorable (a function of temperature and pressure). This is called sublimation.

b. The common use in snow hydrology is for phase changes of water as it condenses from vapor to liquid, or as it melts from solid to liquid form. The latent heat quantities for these phase changes are given in Table C-4.

Table C-4
Latent Heats for Water

	kJ/kg	cal/g
Melt (fusion)	333.5	79.5
Condensation (vaporization), 0 °C	2500	597.3
Condensation (vaporization), 10 °C	2477	591.7
Condensation (vaporization), 20 °C	2453	586.0
Sublimation, 0 °C	2834	677.0
Sublimation, -30°C	2839	678.2

C-5. Heat Transfer

a. Conduction. Transfer of heat within a solid body by molecular activity because of a differential in temperature in the body is called heat conduction. This phenomenon is encountered in snow hydrology when snowmelt caused by heat conducted from the ground is considered. The measure of a material's ability to conduct heat is given by its coefficient of thermal conductivity, k. Thus, heat transferred is given by

$$Q = kA\frac{dT}{dx}$$

where

Q = heat flux

k = coefficient of thermal conductivity, commonly expressed as kW/(m K)

A = cross-section area

dT/dx = temperature gradient

Table C-5 gives values of k for some common substances

Table C-5
Coefficients of Thermal Conductivity

Substance	kW/(m·K)
Ice	2.3×10^{-3}
Limestone	2.2×10^{-3}
Peat	0.08×10^{-3}
Silt and clay	$0.4\text{-}2.1 \times 10^{-3}$
Sandy soils	$0.25\text{-}3 \times 10^{-3}$
Wood	$0.15\text{-}0.20 \times 10^{-3}$
Air	2.3×10^{-5}

b. Convection. This term is applied to heat transfer in a fluid through the movement of the fluid, brought about by natural or induced pressure or density differences. An example of natural convection would be the overturning of a lake as cold air cools the surface layer of the water. In snow hydrology, convection heat transfer is one of the processes by which heat is transmitted through the air to the snow surface. In this case, the air movement is induced by the wind, and the mathematical representation of the phenomenon is based upon equations for turbulent exchange.

c. Radiation. Radiation energy exchange refers to the continual emission of energy, which occurs from all bodies, in the form of electromagnetic waves. When they fall on a body that is not transparent to them, they are absorbed and their energy is converted to heat. The radiant energy emitted by the surface depends upon the nature of the surface and on its temperature. At low temperatures, the rate of radiation is small, but as the temperature of the emitter is increased, the radiation intensity increases very rapidly, in proportion to the 4th power of the absolute temperature of the body. The maximum amount of radiation for a given temperature is called the blackbody radiation. A body that radiates at the maximum intensity for every wavelength at the given temperature is called a blackbody. This term applies regardless of the literal color of the body; the sun is a perfect blackbody, and the surface of snow is nearly so.

(1) The radiant energy is emitted in a mixture of different wavelengths, which can be expressed in the form of a continuous spectral distribution. As the temperature of the emitter increases, there is a general decrease in the wavelength of the maximum intensity. A general equation expressing the energy emitted by a blackbody to the wavelength and temperature was derived by Max Planck in 1900. The total blackbody radiation over all wavelengths at a given temperature is measured by the area under a Planck curve for that temperature. This integration, known as the Stefan-Boltzmann law, yields the equation

$$E = \sigma T^4$$

where σ, the Stefan-Boltzmann constant $= 5.7 \times 10^{-11}$ $kJ/(m^2\ s\ K^4)$

(2) This relationship is used directly in equations for snowmelt, both attributable to solar radiation and from the surface of the snow as a long-wave radiation. Discussion of solar radiation is continued further in Appendix D.

Appendix D
Meteorological Relationships

D-1. Water Vapor in Air

a. Vapor pressure and saturation. Water vapor present in the atmosphere is measured in terms of the partial pressure exerted by the gas, known as the vapor pressure. As the amount of vapor increases for a given temperature, the pressure increases until it reaches a state of equilibrium with a liquid water surface at that temperature. This is called the saturation vapor pressure. Saturation vapor pressure is specifically related to temperature, as shown in Table D-1. Vapor pressures are commonly measured in terms of millibars of pressure.

Table D1
Saturation Vapor Pressure (mb) Over Water and Over Ice (after Byers 1974)

Temperature, °C	Over Water	Over Ice
-10	2.863	2.597
-5	4.215	4.015
0	6.108	6.108
5	8.719	
10	12.272	
15	17.044	
20	23.373	
25	31.671	
30	42.430	
35	56.236	

b. Relative humidity. The relative humidity is defined as the ratio of the measured water vapor content of the air at a specified temperature to the saturated vapor content at that temperature. It can be computed by the ratio of vapor pressures:

$$RH = \frac{e_a}{e_s} \times 100$$

where

RH = relative humidity, percent

e_a = vapor pressure of the air

e_s = saturated vapor at the temperature of the air

Relative humidity is measured by a sling psychrometer, which contains two thermometers, one in which the bulb is covered with a cloth wetted with distilled water. The dry bulb will indicate the air temperature, and the wet bulb will be cooled below the air temperature by evaporation. The amount of evaporation will depend upon how saturated the air is. Tables are available to relate the difference—the wet bulb depression—to relative humidity.

c. Dew point. The temperature at which the air must be cooled to become saturated is called the dew-point temperature. Since the temperature of the dew point is related to vapor pressure, it is used as a surrogate for vapor pressure in snowmelt equations. Dew point can be computed from relative humidity and air temperature as shown in Figure D-1.

D-2. Solar Radiation

a. Solar constant. The solar constant is defined as the rate of radiant solar energy flux received outside the Earth's atmosphere on a surface normal to the Sun's rays. At the mean distance from the Sun, this value is 1.35 kW/m^2, or 1.94 cal/(cm^2 min) (1.94 ly/min). This value varies about 7 percent during the year primarily because of the changing distance between the Earth and Sun.

b. Incident radiation. The spectral distribution (Planck Curve) of the theoretical radiation emitted by the sun is shown in Figure D-2. Solar radiation (shortwave) radiation generally encompasses the wavelength range of 0.2 to 2.2 μm. Radiation emitted by the atmosphere and Earth (long-wave radiation) has a wavelength range of 6.8 to 100 μm

(1) Solar radiation received at the Earth's surface is actually made up of both direct solar radiation, plus a

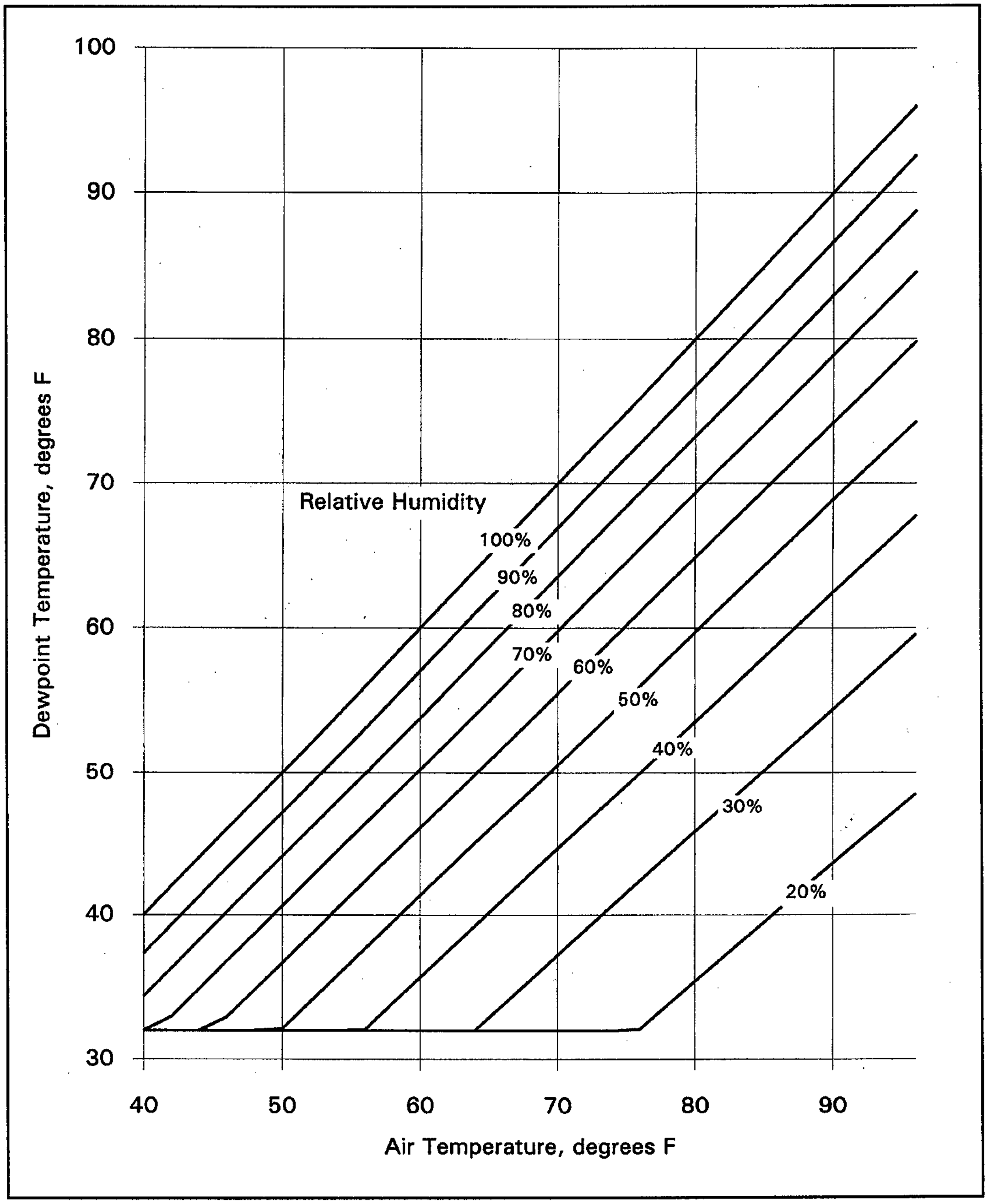


Figure D-1. Dew-point temperature as a function of air temperature and relative humidity

mall component that is scattered by the atmosphere (diffuse or sky radiation). The rate at which the total is received on a horizontal surface is termed insolation. This is expressed as a flux per unit area (flux density), such as watts per square meter or megaJoules per square meter per day. An older convention, used in *Snow Hydrology*, is g-cal/(cm^2 min), or langleys (ly) per minute, where a langley is equivalent to 1 g-cal/cm^2. Another term used to express flux density is irradiance. Table D-2 summarizes the comparisons

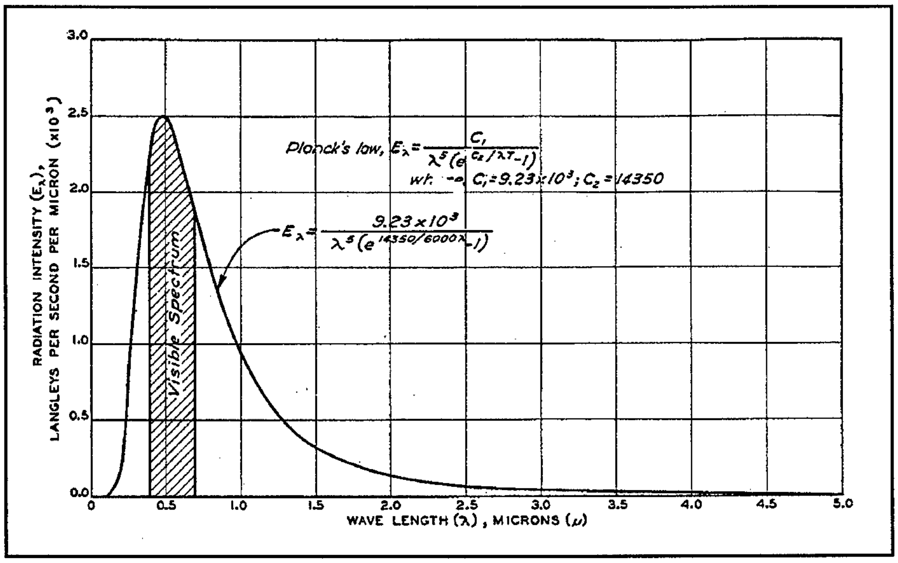


Figure D-2. Spectral distribution (Planck Curve) of the Sun's radiation (Figure 2, Plate 5-1, *Snow Hydrology*)

among three common conventions for expressing insolation. Table D-3 contains typical values of daily insolation at 45° north latitude for conditions outside the atmosphere, and for the Earth's surface assuming a cloudless sky at the maximum (spring equinox) and minimum (winter equinox) sun angles.

Table D2
Conversion Factors for Insolation Units

	ly/day cal/(cm^2·day)	mJ/(m^2 day)	W/m^2
1 ly/day =	1	0.04186	0.4844
1 mJ/(m^2·day)=	23.89	1	11.57
1 W/m^2 =	2.064	0.0864	1

Table D-3
Typical Daily Insolation Values

For Latitude 45° N	Langleys	mJ/m^2	W/m^2
Top of atmosphere, 21 June	990	41	480
Top of atmosphere, 20 Dec	250	11	120
Earth's surface,[1] 21 June	750	31	360
Earth's surface, 20 Dec	200	8	97

[1] For a horizontal surface and a clear day.

(2) Insolation magnitude depends upon the solar constant, the angle of the Sun's rays (a function of season and latitude), and the amount of depletion in the atmosphere. Depletion results from absorption by gas molecules, dust, smoke, etc., and cloud particles. Clouds have by far the greatest effect in reducing the amount of radiation energy received on Earth. Figure D-3 shows the daily insolation amounts outside of the atmosphere, before attenuation by the atmosphere. The effect of atmospheric influences under cloudless skies is shown on Figure D-4, which is based upon measurements at the Central Sierra Snow Laboratory.

(3) The effect of clouds on solar radiation received can be quite pronounced and highly variable. Two factors, the amount of cloud cover (percent of sky covered) and the cloud height, are involved. Figure D-5 illustrates the effect of cloud height and cover.

(4) Another determinant for solar radiation falling upon a surface is the slope of the surface itself. In the northern hemisphere, it is obvious that a south-facing slope will receive more solar radiation than a north-facing slope of the same magnitude. This effect is more pronounced in the winter. Figure D-6 illustrates the effect of slope on incident solar radiation for latitude 46° 30' N.

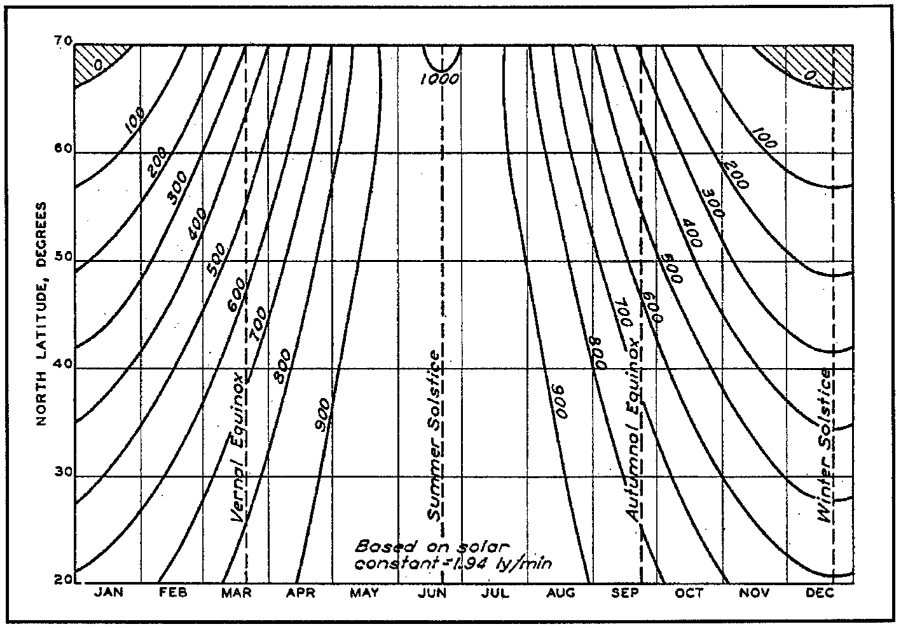


Figure D-3. Seasonal and latitudinal variation of solar radiation outside the Earth's atmosphere (Figure 3, Plate 5-1, *Snow Hydrology*)

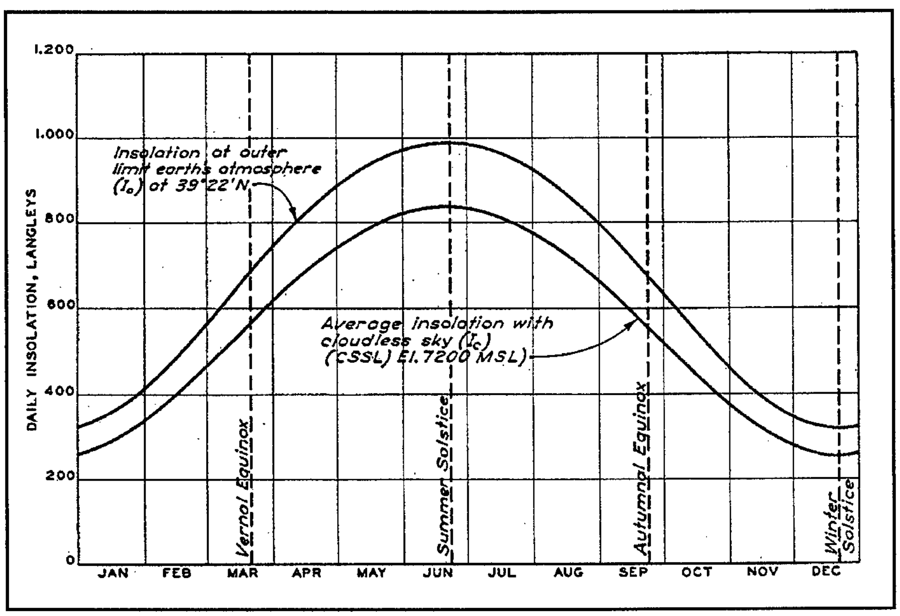


Figure D-4. Seasonal variation in insolation at the Central Sierra Snow Laboratory, showing atmospheric depletion (Figure 4, Plate 5-1, *Snow Hydrology*)

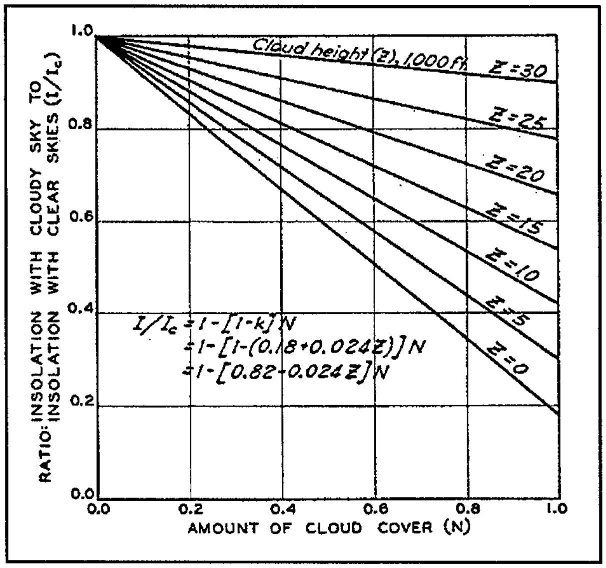


Figure D-5. Variation of insolation with cloud height and amount of cloud cover (Figure 5, Plate 5-1, *Snow Hydrology*)

(5) Forest cover also plays an important part in the amount of solar energy that reaches the snow surface. For only coniferous forests, the transmission percentage varies with the season, because of variation in the shading effect of the trees with the solar altitude. The determination of the amount of sunshine transmitted through the forest is at best approximate. Figure D-7 shows a mean transmission curve for daily insolation amounts, expressed in terms of forest canopy density. In the generalized snowmelt equations, the transmission coefficient and forest density are combined into a single factor *F*, which is termed the effective forest cover.

(6) One way of expressing the effect of cloud cover is in terms of percentage of possible sunshine. With this as a variable, a practical nomograph has been developed to estimate daily insolation as a function of latitude and season. This is shown in Figure D-8.

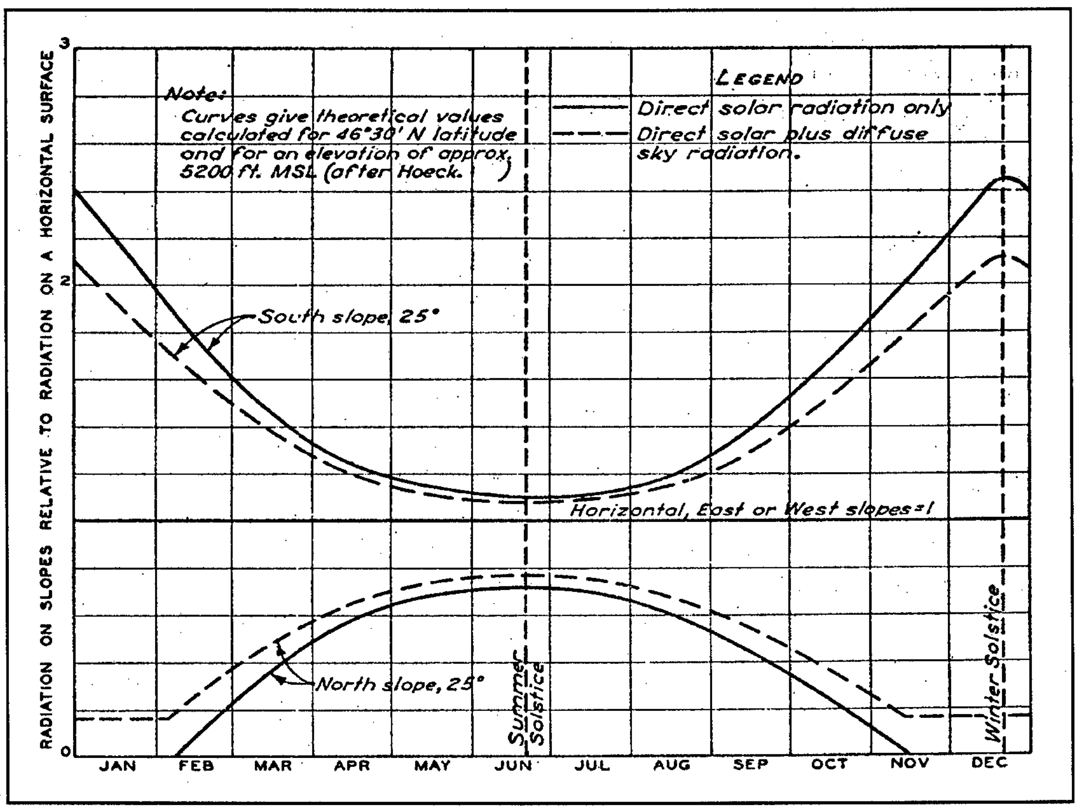


Figure D-6. Seasonal variation— radiation on slopes versus radiation on a horizontal surface (Figure 5, Plate 5-1, *Snow Hydrology*)

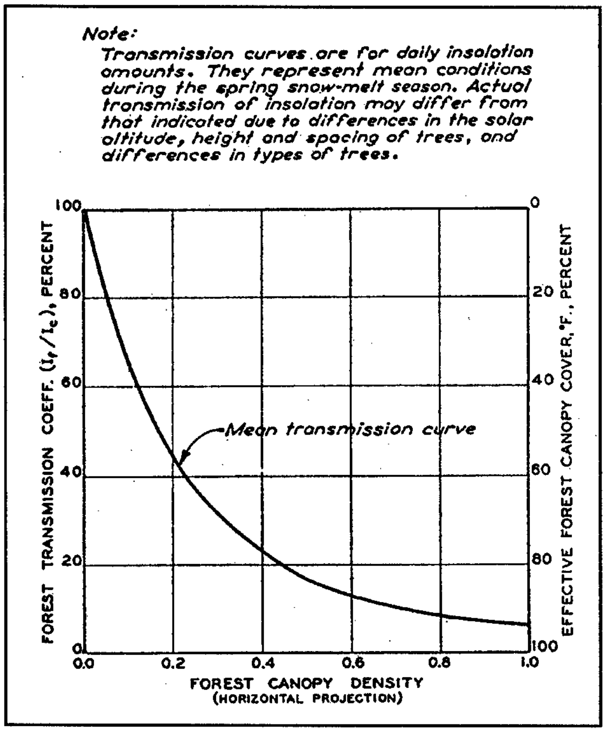


Figure D-7. Transmission of solar energy by a forest canopy (Figure 1, Plate 5-2, *Snow Hydrology*)

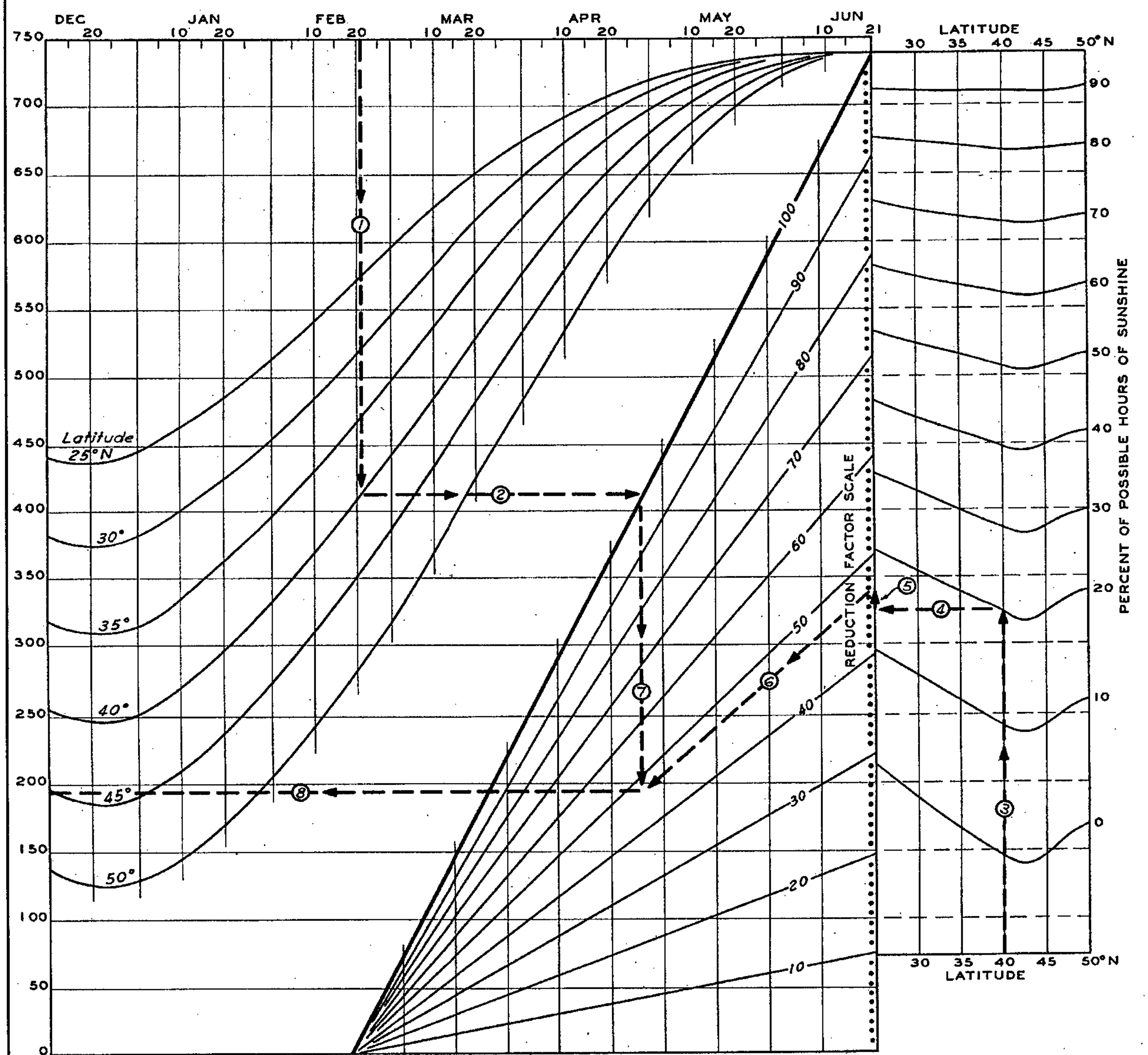
D-3. Long-wave Radiation

Long-wave or thermal radiation, emitted by the sky and Earth, encompasses wavelengths from about 6.8 to 50 μm. Figure D-9 is the spectral distribution of radiation intensity for a blackbody at 0 °C, which is approximately equivalent to melting snow. Since snow is nearly a blackbody, the outgoing long-wave radiation is essentially a constant, computed by Stephen's law as 0.45 ly/min. Back-radiation (towards the Earth) is emitted by the atmosphere, clouds, and forest cover and is a complex phenomenon that must be computed experimentally. The net long-wave radiation is equal to outgoing radiation flux less back-radiation.

a. Net radiation from clear skies. Radiation from the atmosphere can be expressed in terms of the temperature of the air and its moisture content, the latter measured by vapor pressure of the air. Figure D-10, based upon experimental evidence, illustrates the net radiation associated with open clear skies. This shows that most of the time there is an outgoing flux of radiation under clear skies—the air temperature must be 69 °F for a gain to the snowpack to occur.

b. Net radiation with cloud cover. Figure D-11 is a curve representing the theoretical net exchange under overcast skies, which are assumed to be radiating as a blackbody. This curve further illustrates the effect of cloudy skies in reducing the radiation loss that would occur for the same temperature under clear skies.

c. Net radiation with forest cover. The presence of a forest canopy is a somewhat similar situation to that of cloud cover with regard to net radiation exchange with the snowpack. The canopy, if a solid cover, absorbs and emits all possible radiation, acting at the temperature of the tree leaves, which is approximately the ambient air temperature. This effect is illustrated in Figure D-12.



Notes:

1. The sample shown by dashed lines with arrows estimates the daily total insolation at a station at latitude 40°N, on February 21 with 20 percent possible sunshine: Step ① – enter the graph at February 21 and move downward to the curve labelled 40°N. Step ② – thence move horizontally to establish a point at the intersection with the heavy line. Step ③ – enter lower right of graph at 40°N, proceeding upward to 20 percent of possible hours of sunshine. Step ④ – thence move horizontally toward the left to the vertical line through 25°N. Step ⑤ – here the seasonal correction (+2) from the table at the right is added on the reduction factor scale. Step ⑥ – thence move downward toward the convergence point. Step ⑦ – from the reference point of step 2 move downward to an intersection with the line extended in step 6. Step ⑧ – from the point of intersection of these two lines move horizontally to the insolation scale to read the estimated value, 195 langleys per day.
2. For use between June 21 and December 21, the curves are symetrical about June 21.

SEASONAL CORRECTION TO REDUCTION FACTOR

Month	Percent Of Possible Sunshine							
	0	10	20	30	40	50	60	70
Jan	+4	+3	+3	+2	+2	+2	+1	+1
Feb	+3	+3	+2	+2	+2	+1	+1	+1
Mar	−1	−1	−1	−1	−1	0	0	0
Apr	−2	−2	−1	−1	−1	−1	−1	0
May	−4	−3	−3	−2	−2	−2	−1	−1
Jun	−5	−4	−4	−3	−2	−2	−2	−1
Jul	−5	−4	−3	−3	−2	−2	−2	−1
Aug	−4	−3	−3	−2	−2	−2	−1	−1
Sep	−2	−2	−1	−1	−1	−1	−1	−1
Oct	0	0	0	0	0	0	0	0
Nov	+2	+2	+1	+1	+1	+1	+1	0
Dec	+4	+3	+3	+2	+2	+2	+1	+1

Figure D-8. Nomograph for estimating insolation as a function of latitude, date, and duration of sunshine (Figure 3, Plate 6-1, *Snow Hydrology*)

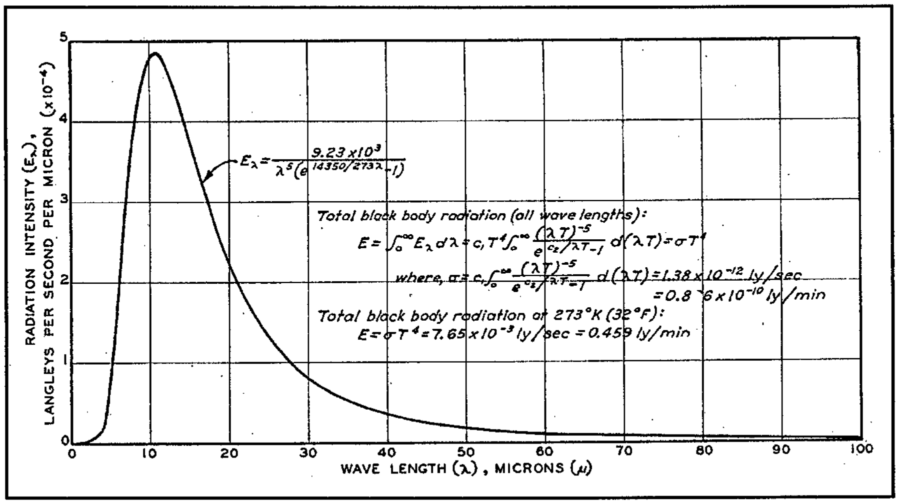


Figure D-9. Theoretical spectral distribution for a snow surface at 0 °C (Figure 1, Plate 5-3, *Snow Hydrology*)

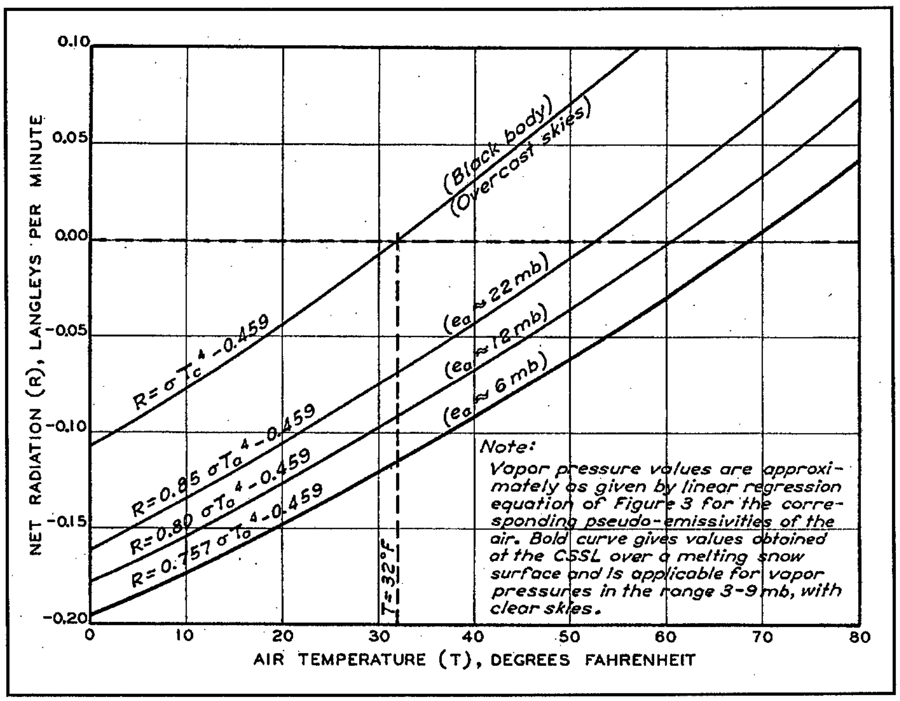


Figure D-10. Net long-wave radiation exchange between the snowpack and the atmosphere, clear skies (Figure 4, Plate 5-3, *Snow Hydrology*)

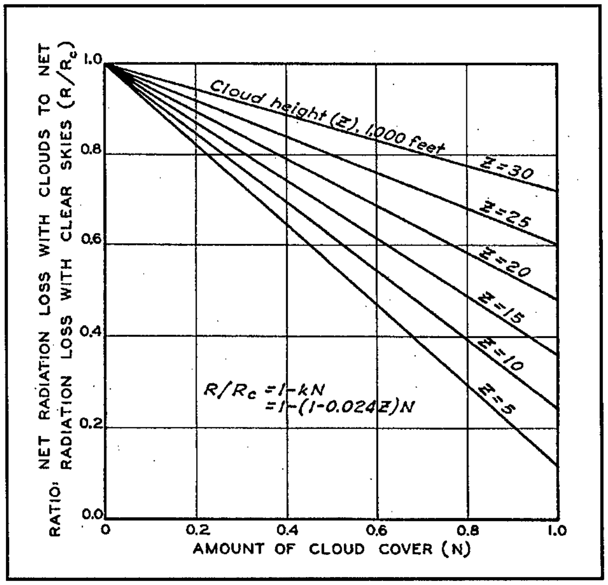


Figure D-11. Variation in net long-wave radiation loss with cloud height and amount (Figure 5, Plate 5-3, *Snow Hydrology*)

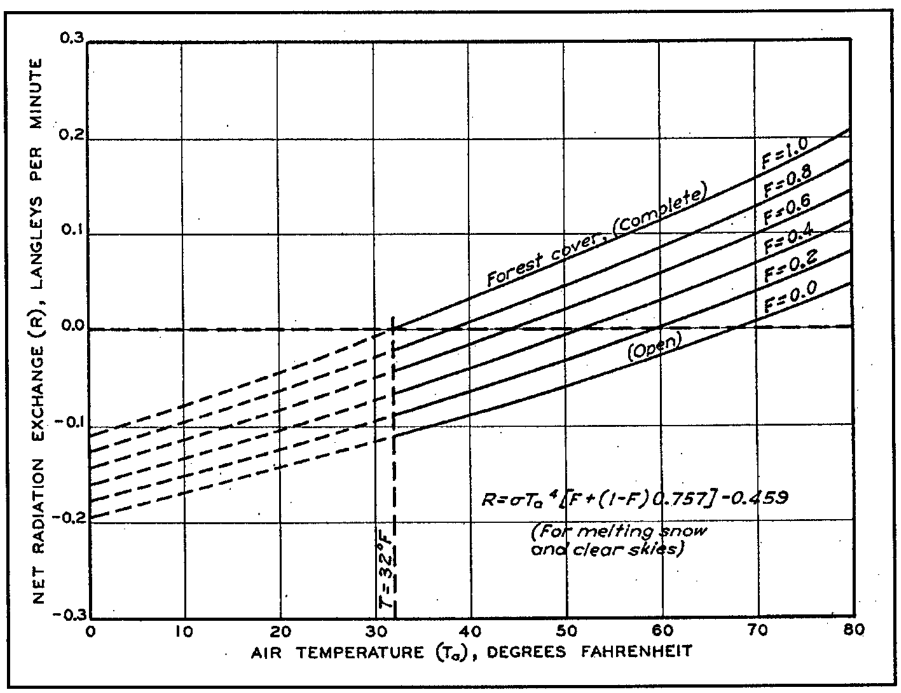


Figure D-12. Net long-wave radiation exchange in forested areas (Figure 6, Plate 5-3, *Snow Hydrology*)

Appendix E
Metric (SI) Versions of Generalized Snowmelt Equations

E-1. General

Reference should be made to Chapter 5 of the main text for an explanation of the coefficients and for the background on the equation derivations.

E-2. Snowmelt During Rain

a. Partly forested areas.

$$M = (1.33 + 0.239\,kv + 0.0126P_r)T_a + 2.3$$

b. Forested areas.

$$M = (3.38 + 0.0126P_r)T_a + 1.3$$

where

M = snowmelt, mm/day

k = basin wind coefficient

v = wind velocity at the 15-m height, km/hour

P_r = daily rainfall, mm

T_a = mean temperature of the saturated air, °C

E-3. Rain-Free Snowmelt

a. Open areas.

$$M = k'(1\text{-}F)(3.08\,I_i)(1\text{-}a) + (1\text{-}N)(0.969T'_a - 21.34) + N(1.33T'_c) + k(0.239v)(0.22T'_a + 0.78T'_d)$$

b. Partly forested areas.

$$M = k'(1\text{-}F)(2.43\,I_i)(1\text{-}a) + k(0.239v)(0.22T'_a + 0.78T'_d) + F(1.33T'_a)$$

c. Forested areas.

$$M = k(0.239v)(0.22T'_a + 0.78T'_d) + F(1.33T'_a)$$

d. Heavily forested areas.

$$M = 3.38(0.53T'_a + 0.47T'_d)$$

In the above equations

M = snowmelt, mm/day

k' = basin shortwave radiation melt factor

F = average basin forest canopy cover, effective in shading the area from solar radiation, expressed as a decimal fraction

I_i = insolation (solar radiation on horizontal surface), MJ/m^2

a = average snow surface albedo

N = estimated cloud cover expressed as a decimal fraction

T'_a = difference between the air temperature measured at 3 m and the snow surface temperature, °C

T'_c = difference between the cloud base temperature and snow surface temperature, °C

k = basin convection-condensation melt factor

T'_d = difference between the dew-point temperature measured at 3 m and the snow surface temperature, °C

v = wind velocity at the 15-m height, km/hour

Appendix F
Summary Descriptions of Selected Operational Snowmelt Models

F-1. General Introduction

Many models have been created around the world over the last four decades or so to describe snowmelt runoff. Some 18 different models are listed and summarized in the *Snow Hydrology Guide* (Ontario Ministry of Natural Resources 1989). The World Meteorological Organization (1986) also lists and summarizes 18 different snowmelt runoff models. These many models are listed here in Table F-1.

a. This section will focus on describing six snowmelt models that have been demonstrated as valuable operational models or are thought to have a high potential for future operational use by the U.S. Army Corps of Engineers. These models are as follows:

- The SSARR Model (the Streamflow Regulation and Reservoir Regulation Model).
- The HEC-1 and HEC-1F Models (the Hydrologic Engineering Center - 1, 1F Model).
- The NWSRFS Model (the National Weather Service River Forecast System Model).
- The PRMS Model (the Precipitation Runoff Modeling System Model)
- The SRM (the Snowmelt Runoff Model).
- The GAWSER Model (the Guelph All-Weather Storm-Event Runoff Model).

b. In the following sections, the theoretical basis and application of each of these six models will be briefly described as will their data requirements and significant features. Each model description will include important citations relating to model development and use.

F-2. Brief Descriptions of Snowmelt Models

a. SSARR model. The Streamflow Synthesis and Reservoir Regulation model was originally developed by the North Pacific Division of the U.S. Army Corps of Engineers in 1956 (USACE 1956). This model has been successfully applied to numerous river systems as diverse as the Columbia and Mekong rivers (Rockwood 1978) and is well documented within USACE (1991). Viessman et al. (1977) state that SSARR is one of the earliest continuous streamflow simulation models using lumped parameter representation and has its primary strength in its verified accuracy.

(1) The conceptual logic underlying the watershed model (SSARR also has river system and reservoir system models) is shown schematically in Figure F-1. SSARR watershed model can be visualized as comprising two modules, the snow computation module and the runoff analysis module. The Runoff Analysis Module uses a single soil-moisture reservoir whose level or state determines the percentage of available precipitation or snowmelt that eventually runs off via combined surface, subsurface, and base-flow components. Water that does not run off is apportioned between soil-moisture reservoir gains and evapotranspiration losses. At present the operational SSARR model does not deal directly with moisture of frozen ground or the temperature-dependence of important water properties that affect runoff.

(2) Within the snow computation module, the SSARR program computes snowmelt through the use of two options that allow it to be tailored to specific applications. The first option for computing snowmelt is based on a temperature index approach, while the second option is the generalized snowmelt equation as derived from *Snow Hydrology* (USACE 1956). Within this module, the state of the basin snowpack can also be defined by two different options: the snowcover "depletion curve" option or the "integrated-snowband" option.

(3) The depletion curve model computes snowmelt with an algorithm that is based on the

Table F-1
Listing of Snowmelt Models
(As identified by the Ontario Ministry of Natural Resources (1989) and the World Meteorological Organization (1986))

Model Name	Country of Origin	Reference
Point Energy/Mass Balance Model	USA	Anderson (1976)
HSP-F (Hydrologic Simulation Program-Fortran)	USA	Johanson et al. (1984)
NWSRFS (National Weather Service River Forecast System)	USA	Anderson (1973)
SSARR (Streamflow Simulation and Reservoir Regulation)	USA	USACE (1991)
HEC-1 (Hydrologic Engineering Center-1)	USA	USACE (1990)
USDAHL-74 (Revised Model of Watershed Hydrology)	USA	WMO (1986)
SCS (SCS Snowmelt Model)	USA	WMO (1986)
SWMM (Storm Water Management Model)	USA	WMO (1986)
USGS (U.S. Geological Survey Model)	USA	WMO (1986)
SIMFLO (Continuous Streamflow Simulation Model)	Canada	Bishop and Watt (1975)
GAWSER (Guelph Agricultural Watershed Storm-Event Runoff Model)	Canada	Ghate and Whiteley (1977)
MOEHYDRO2 (Comprehensive Watershed Model)	Canada	Logan (1976)
WRB (Water Resources Branch Model)	Canada	Kite (1978)
UBC (University of British Columbia Watershed Model)	Canada	Quick and Pipes (1977)
QFORECAST (Continuous Simulation and Real-Time Forecast Model)	Canada	WMO (1986)
SRM (Snowmelt Runoff Model)	Switzerland	Martinec (1975)
HBV (Conceptual Runoff Model for Swedish Catchments)	Sweden	Bergström (1975)
SHE (Systems Hydrologique European Snow Model)	France	Morris and Godfrey (1978)
CEQUEAU	Canada	WMO (1986)
ERM (Empirical Regressive Model)	Czechoslovakia	WMO (1986)
NEDBOR-AFSSSTROMNINGSMODEL (Rainfall -Runoff Model v. II)	Denmark	WMO (1986)
TANK (Tank Model with Snow Model)	Japan	WMO (1986)
IHDM (Institute of Hydrology Distributed Model)	UK	Morris (1983)
PRMS (Precipitation-Runoff Modeling System)	USA	Leavesley et al. (1983)
YETI	Czechoslovakia	WMO (1986)
SCHNEE	GDR	WMO (1986)
WSRM (Winter Season Runoff Model)	Poland	WMO (1986)
HRO (Hydro Resources Optimization)	USA	WMO (1986)
GMTs-1 (Model of Snowmelt Formation of Lowland Rivers)	USSR	WMO (1986)
GMTs-2 (Model of Snowmelt Formation in a Mountainous Basin)	USSR	WMO (1986)
GMTs-3 (Model of Snowmelt-Rainfall Runoff Formation)	USSR	WMO (1986)

temperature index or energy budget and a snow cover depletion curve. The depletion curve is based on a theoretical relationship between a snow-covered area as a percentage of watershed area versus accumulated generated runoff as a percentage of seasonal total. The actual snow-covered area and accumulated runoff for a computation period are compared with the theoretical snowcover. This approach can treat a watershed as a single lumped unit or as a split watershed with snow-covered and snow-free areas, as in the case of mountainous watersheds where there is a snowline.

(4) The integrated-snowband provides the ability to formulate the watershed into bands of equal elevation, on which snow accumulation and ablation, as well as soil moisture, are accounted for independently. Key elements include the following:

- Snow conditioning or accounting for the snowpack heat deficit.

- A vegetation interception algorithm.

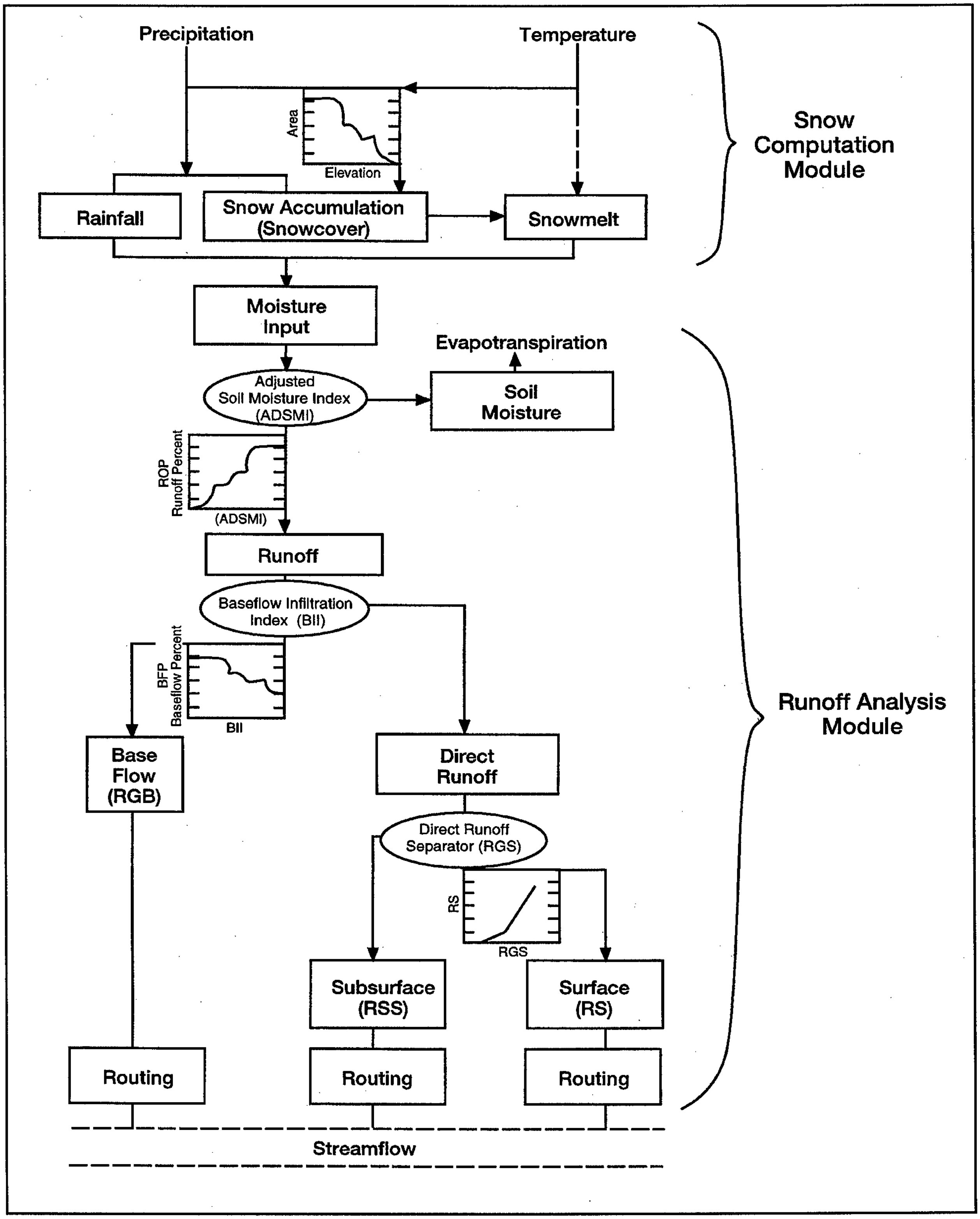


Figure F-1. Flowchart of SSARR Model

- A flexible evaporation simulation.
- Routing to simulate long-term return flow from groundwater.

Snowmelt is calculated using the temperature index method during a nonrain event and by a modified melt equation for snowmelt during a rain event in a heavily forested area. The integrated-snowband model uses Anderson's (1978) heat deficit approach for its snowpack conditioning routine. Liquid water does not enter the soil system to be available for runoff until the "cold content" and snowpack liquid water deficiency are satisfied. Ground melt resulting from conduction of heat from the ground is assumed to be constant or a function of the month of the year. This logic is summarized in Figure F-2.

(5) The SSARR model program is written in IBM-VS FORTRAN-77. It has also been made available for the VAX-11/780 computer and IBM PC-compatible microcomputers. Data management and analysis programs to support operational day-to-day forecasting and long-term simulations are also available (USACE 1991), and interface with HECDSS is possible. Data are input in fixed-column card formats, free-form card formats, or as responses to prompting messages by an interactive driver. Output has a wide range of formats and varies from plots of key variables and statistics to "card-image" output that may be used for subsequent SSARR runs.

b. HEC-1 and HEC -1F models. The HEC-1 Flood Hydrograph Package is a flood runoff event simulation model first developed in 1967 by the Hydrologic Engineering Center of the USACE. It has been revised and updated a number of times to improve its computational methods and user interface (USACE 1990). It has also been connected to the HEC Data Storage System (DSS) for storage and retrieval of data and improved graphical and tabular output capabilities. HEC-1 is a generalized program that simulates the runoff from snowmelt or rainfall, or both, for virtually any type of watershed or river basin. There is no limit to the size or number of subbasins and routing reaches needed to describe a basin. The HEC-1F program is a special version of HEC-1 for use in real-time forecasting. It includes real-time optimization and blending routines. Some HEC-1 options are not available in HEC-1F, however.

(1) HEC-1 is basically a general calling program that can access any one of a number of options within six subroutines. These subroutines are as follows:

- Optimal determination of unit hydrographs.
- Streamflow routing.
- Snowmelt computations.
- Unit hydrograph computations.
- Hydrograph routing and combining computations.
- Hydrograph balancing computations.

In addition to the basic hydrological simulation, HEC-1 has several capabilities to assist in hydrological investigations. These capabilities include the following:

- Automated parameter estimation for Infiltration Rates, Unit Hydrographs and Streamflow Routing.
- Snowmelt parameter estimation.
- Dam breach simulation.
- Automatic precipitation depth area adjustments.
- Multiple basin developments and storm size simulation.
- Streamflow diversions and pumping plants.
- Flood damage compilation.
- Flood frequency curve modification.
- Annual flood damage expectation.
- Flood control projects size optimization.

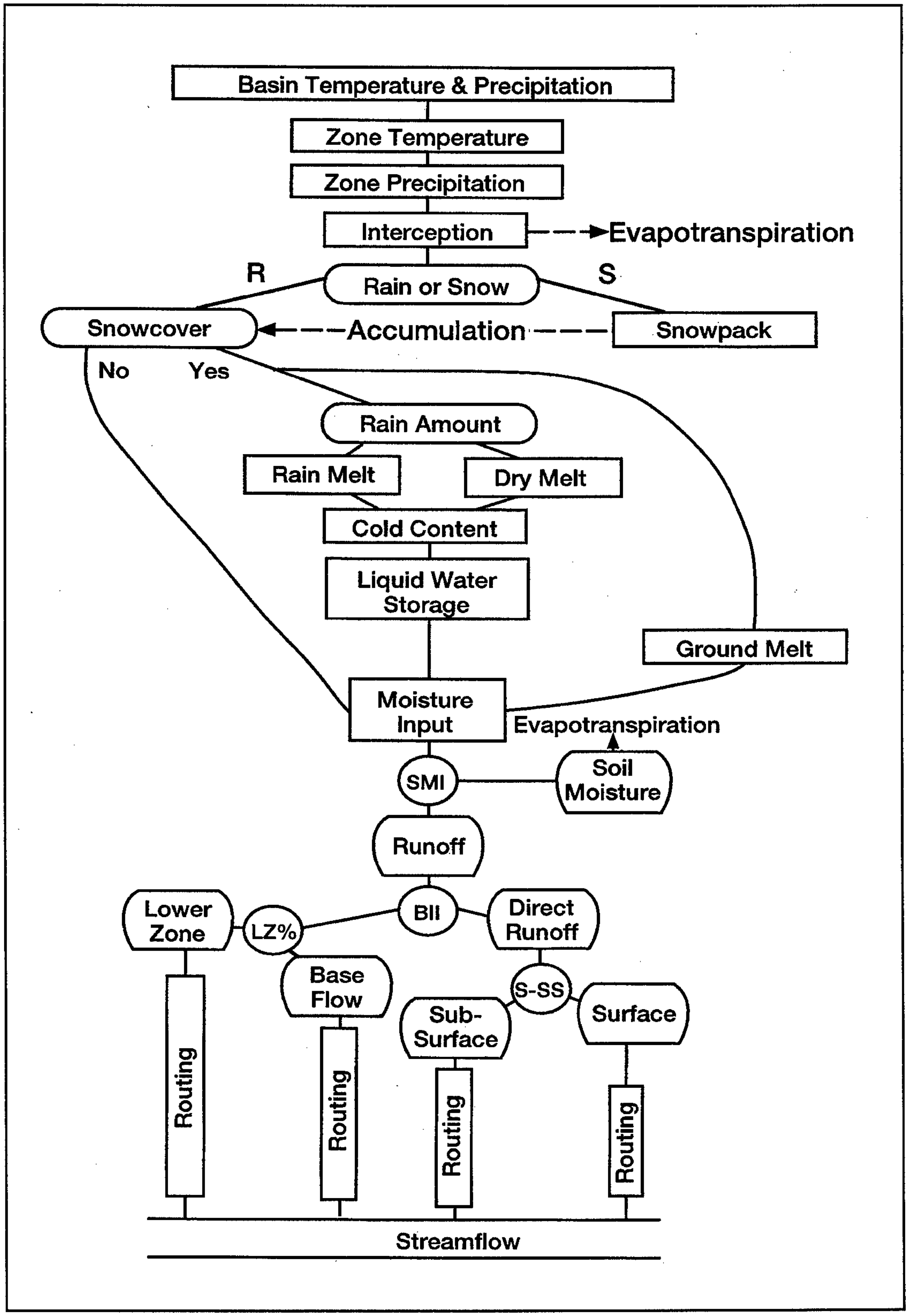


Figure F-2. SSARR Integrated Snowband Model

(2) HEC-1 is an event-type model, applicable for modeling flood runoff only. Runoff is simulated by applying rainfall and snowmelt to a unit hydrograph, then computing the total hydrograph by adding base flow. Several loss-rate functions are available. There is no representation of the effects of frozen ground. There is no direct accounting for water properties that change with temperature.

(3) Snowmelt is calculated using either the degree-day (temperature index) or energy budget methods as described in *Snow Hydrology* (USACE 1956). The energy budget approach is used for rain-on-snow events. There is a provision to account for up to 10 elevation zones within a subbasin, with the temperature being lapsed in degrees per increment of elevation in each zone. Snow accumulation is accounted for and precipitation may fall as rain or snow, depending on zone temperature. Heat deficit or the "ripeness" of the pack are not considered.

(4) HEC-1, including the DSS interface, is written in ANSI standard FORTRAN 77 as is available on the IBM PC, mainframe, and UNIX-based workstation computers (USACE 1990), and on the Macintosh computing platform. Since DSS offers a wide range of input and output options, as well as access to many databases that are necessary for modeling large-scale river systems, the DSS interface with HEC-1 is an important feature for the operational use of this model.

c. NWSRFS model. The National Weather Service River Forecast System (NWSRFS) model is a further development of the Standard Watershed Model (Crawford and Linsley 1966). It was developed in 1972 by the Hydrologic Research Laboratory (HRL) of the NWS, Office of Hydrology. The Snow Accumulation and Ablation Model within the NWSRFS model is described in HYDRO-17 (Anderson 1973).

(1) The NWSRFS model uses the Sacramento soil-moisture accounting model (Burnash, Ferrall, and Richard 1973), which divides soil moisture among five reservoirs, using both "free" water and "tension" soil-moisture levels. Available runoff is computed also using the Sacramento soil-moisture accounting model and is translated to runoff using a unit hydrograph approach. An index approach for dealing with frozen ground has been implemented (Anderson and Neuman 1984). No temperature effect on water-holding capacities or rounding constants are accounted for.

(2) The snow accumulation and ablation model described in HYDRO-17 is one of the most successful operational applications of air temperature-index methods. As is stated in HYDRO-17, "The basic philosophy of the model is that each significant physical component be represented separately, rather than to use a single index to explain several processes ...." This is accomplished in NWSRFS with only air temperature and precipitation as the necessary meteorological input parameters. A flowchart showing the basis for the snow accumulation and ablation model in NWSRFS is given in Figure F-3.

(3) The snow accumulation and ablation model in NWSRFS includes consideration of the important components of the energy budget of the snowpack, including snowpack accumulation, heat exchange at the air/snow interface, areal extent of snowcover, heat storage within the snowpack, liquid-water retention and transmission, and heat exchange at the ground/snow interface. Snowmelt is calculated differently for rain and no-rain periods. Melt during nonrain periods is calculated using a degree-day approach, employing a seasonally varying melt-factor. Melt during rain is computed from an energy balance equation that calculates the net radiative, latent, sensible, and rainwater heat transfer to calculate the amount of melt. A key feature of NWSRFS is its snow-conditioning accounting that simulates the cold content and liquid water available in the pack and thereby characterizes the "ripeness" of the snowpack. Areal distribution of the snowpack is dealt with using an areal depletion curve that relates extent of the snow cover to the ratio of mean areal snow water equivalent. This areal depletion curve is considered to be constant from year to year for a particular modeled area. In either rain or nonrain cases, once the heat deficit of the snowpack has been satisfied, the available melt water is lagged and attenuated to simulate the transmission of water through snow. The final excess liquid water is then made available to the runoff portion of NWSRFS.

(4) The snow accumulation and ablation model in NWSRFS is written in FORTRAN IV, has typically been run on IBM mainframe computers (IBM 360/195 at NWS River Forecast Centers), and has been widely

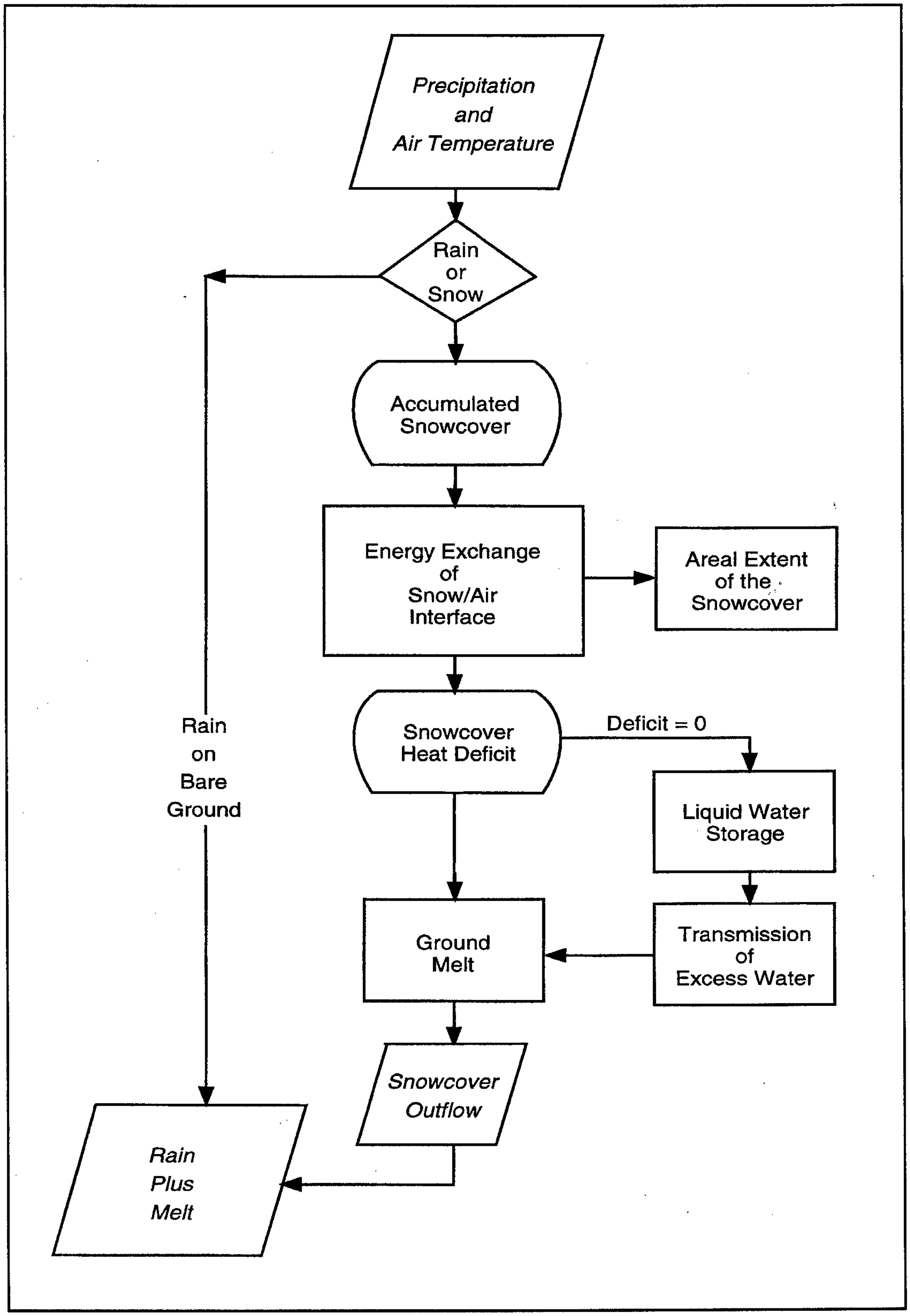


Figure F-3. Flowchart of NWSRFS Snow Accumulation and Ablation Model (after Anderson 1978)

used in research studies. It is fully supported by the Hydrologic Research Laboratory, Office of Hydrology, NWS, and has been used in joint USACE/NWS operational modeling activities (Burnash, Ferrall, and Richard 1973).

d. The PRMS model. The Precipitation-Runoff Modeling System (PRMS) was developed by the U.S. Geological Survey, Water Resources Division, in 1973 (Leavesley 1973). According to Leavesley et al. (1983), PRMS was developed to "evaluate impacts of various combinations of precipitation, climate and land use on surface water runoff...." It is a multipurpose model for stormflow hydrographs and long-term simulations of mean daily runoff from snowmelt. The relationships between available runoff and streamflow are based on a deterministic physical-process model. PRMS is a modular-design modeling system to provide a flexible modeling capability. The PRMS is structured into three major components: the data management component, the PRMS library component, and the output component. These three components are shown schematically in Figure F-4. The model is discussed in detail in Leavesley et al. (1983).

(1) A key feature of PRMS allows it to function as a lumped or distributed parameter type model. PRMS allows the watershed to be disaggregated into subareas called Hydrologic Response Units (HRUs) on the basis of soils, vegetation, and climatic and physiographic characteristics. Each HRU is then modeled with the parameters being lumped within the HRU. With the increased availability of Geographic Information Systems (GIS) to USACE field-operating agencies, the disaggregation of basins into HRUs is becoming more practical.

(2) PRMS must receive input variables that describe the physiography, vegetation, soils, climate, and hydrological characteristics of each HRU. The minimum input parameters for driving this model are daily maximum and minimum temperatures, precipitation, and solar radiation.

(3) Snowmelt is modeled using an energy budget approach, as presented by Obled and Rosse (1977). The snowpack routines account for initiation, accumulation, and depletion of the snowpack for each HRU. The energy budget considers net shortwave and longwave radiation, as well as the heat content of precipitation. The snowpack routine accounts for water equivalent and heat deficit and thereby considers the ripeness of the snowpack. Condensation, advection, and ground conduction are not accounted for in the energy budget terms. Frozen ground or the temperature dependence of important water properties are also not included.

(4) The runoff is computed from each HRU using a series of linear and nonlinear reservoirs whose output sums to stream outflow. These reservoirs depict surface flow, subsurface flow, and base flow. In practice, each HRU has its own surface flow reservoir; however, there is typically only one subsurface and one base-flow reservoir for an entire basin. More individual subsurface reservoirs are used for each HRU, depending on the variability of soil characteristics in the basin. The hydrological responses of the individual HRUs are summed to compute the total watershed runoff. A schematic diagram of the concepts that underlie PRMS is presented in Figure F-5.

(5) PRMS is written in FORTRAN 77 and can be run on any machine with this compiler. The model is fully supported by the USGS, Water Resources Division, Denver, CO, and is documented in Leavesly et al. (1983).

e. SRM model. The Snowmelt Runoff Model (SRM) was originally developed in 1973 at the Federal Institute for Snow and Avalanche Research in Davos, Switzerland (Martinec 1975). The SRM simulates or forecasts daily average streamflow in mountainous basins where snowmelt is a major contributor to runoff (Martinec, Rango, and Major 1983). The model has been applied to watersheds ranging from 2.65 km^2 (1.66 square miles) to 4,000 km^2 (2,500 square miles) in both humid and semiarid climates with no serious limitations (Martinec, Rango, and Major 1983; Rango 1989). It is necessary, however, to carefully define the model parameters and variables if accurate results are to be obtained.

(1) SRM uses percentage areal snowcover, air temperature, and precipitation as critical input variables. SRM divides the watershed into elevation zones and accounts for degree-days in each elevation zone to drive the amount of snowmelt. Specific basin

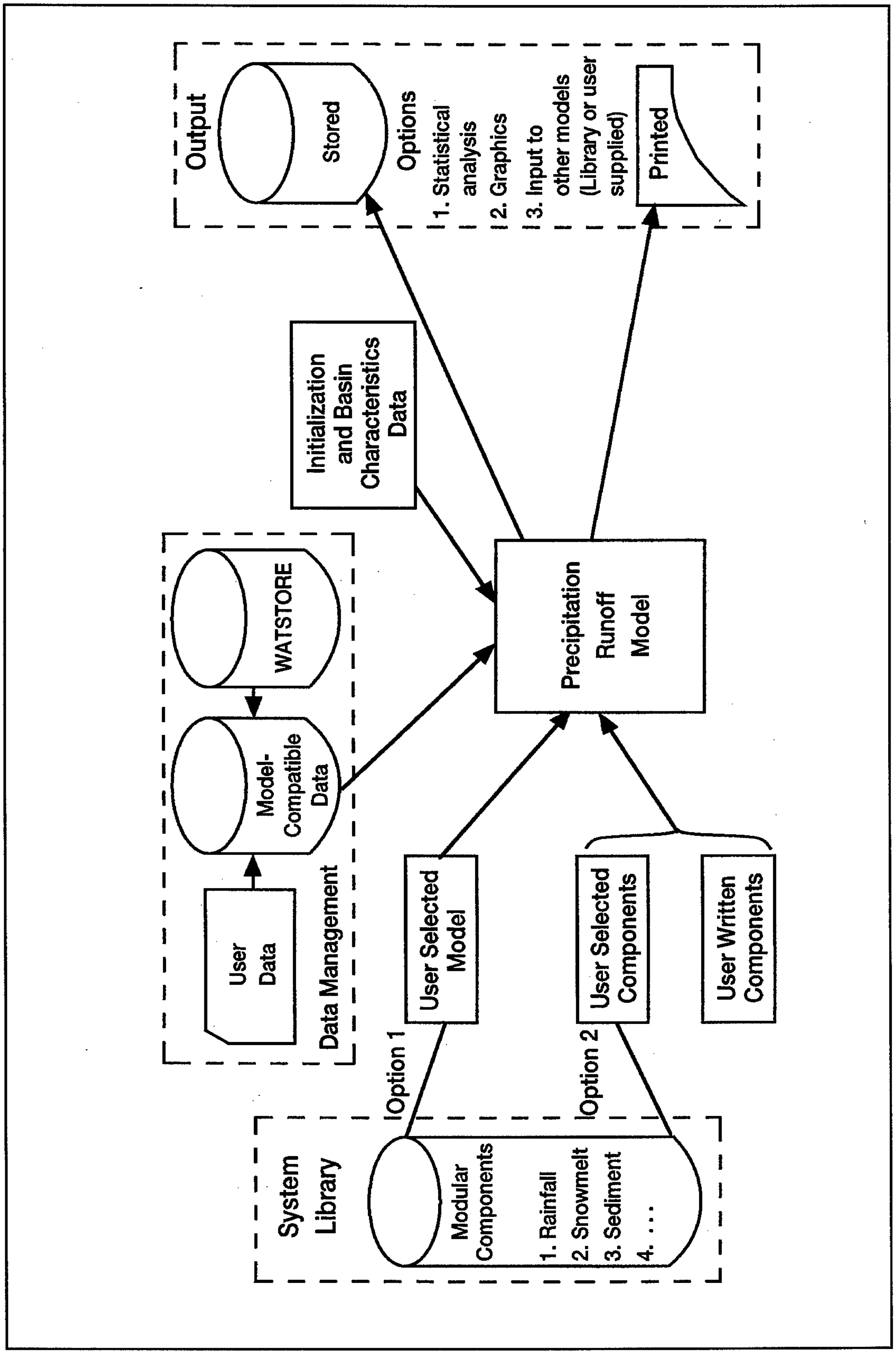


Figure F-4. Flowchart of PRMS (after Leavesley et al. 1983)

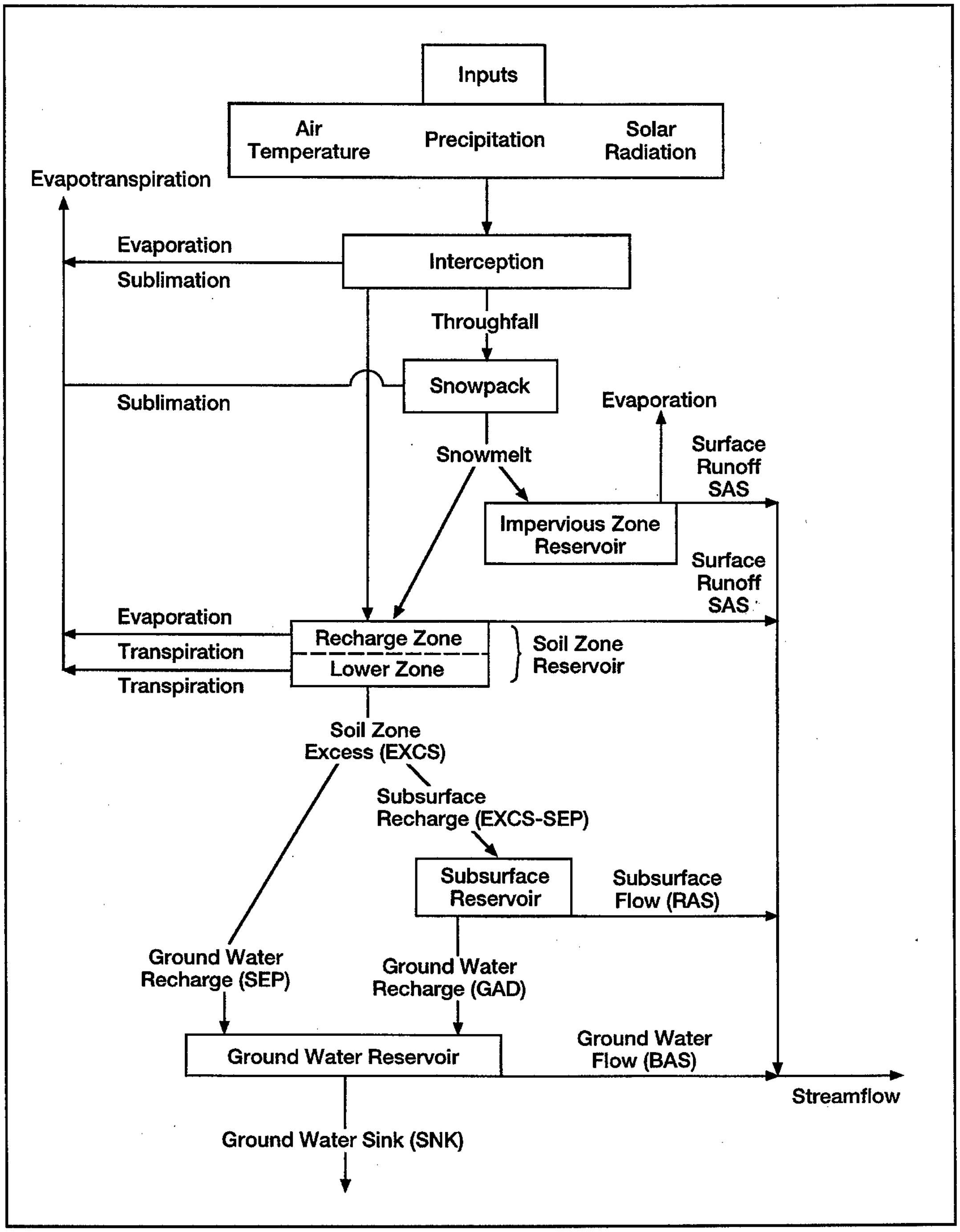


Figure F-5. Schematic of PRMS concepts (after Leavesley et al. 1983)

characteristics include runoff coefficients, degree-day factors, and historical recession coefficients (Shafer, Jones, and Frick 1982). Definition of the basin includes careful determination of basin areas and, once the elevation zones are established, finding the area of each zone. The zonal mean hyposometric elevation is determined for each zone from an area-elevation curve. It is also necessary to know the temperature lapse rate for the basin.

(2) In SRM, "Each day during the snow melt season, the water produced from snow melt and from rainfall is computed, superimposed on the calculated recession flow and transformed into daily discharge from the basin" (Martinec, Rango, and Major 1983). A simple transformation model computes runoff using empirical constants and coefficients for runoff, snowmelt-degree-days, and flow recession. The snowmelt water and precipitation are calculated and superimposed on a calculated recession flow to obtain daily discharge. The strength of SRM is its primary reliance on snow cover areal extent. This allows for limited data input needs, and the snow-covered-area data can be derived from satellite, aircraft, or ground measurements.

(3) Through the use of the zonal mean hyposometric elevations, the actual elevation of the temperature measurement station and the temperature lapse rate, the melting degree-days for each elevation zone are calculated. The precipitation for each zone is determined to be either rain or snow, depending on the average zonal temperature and a critical temperature selected to be slightly above freezing. The snow coverage for each zone is determined by ground observation, aircraft photography, or by satellite and is arrayed as a depletion curve over the snowmelt period.

(4) Runoff coefficient estimation requires knowledge of the basin and its hydrology, and it varies over the year (Martinec, Rango, and Major 1983). The snowmelt-degree-day factor can be varied throughout the snowmelt period to account for the changing density and albedo of the snowpack. The recession coefficient is estimated from historical records of the actual daily average flows.

(5) SRM accumulates the number of degree-days in each elevation zone over the snowmelt period and discriminates the input precipitation into snow or rain by comparing the assigned critical temperature to the average daily temperature. Snowmelt is calculated using a degree-day factor that is applied to the portion of the elevation zone that is snow covered. Within each elevation zone, an average snow cover depletion curve is used to estimate the temporal change in the snow-covered area. The snowmelt is distributed according to the chosen elevation zones and summed to give total average daily runoff from the entire watershed.

(6) SRM is written in FORTRAN and has been documented in Martinec, Rango, and Major (1983). Although originally run on mainframes, the SRM has been modified to a microcomputer version (Rango and Roberts 1987) by the Agricultural Research Service, Beltsville, MD.

f. GAWSER model. The Guelph All-Weather Storm-Event Runoff (GAWSER) model was originally created at the School of Engineering, University of Guelph, in 1977 (Ghate and Whiteley 1977). It is a modification of the HYMO program developed by the U.S. Department of Agriculture in 1972 (Williams and Hahn 1972). Since 1977 it has evolved from a research tool to a fully operated package for synthesis of storm-event hydrographs and large basin reservoir regulation (Grand River Conservation Authority 1989). GAWSER version 5.4 is documented in the GAWSER Training Guide and Reference Manual (GRCA 1989).

(1) GAWSER separates each subwatershed in a basin system into impervious and pervious zones (see Figure F-6). All rainfall and snowmelt incident on impervious zones are routed to overland flow. The pervious areas are desegregated into four soil types (or fewer), with each type being modeled as a two-layered system. Available rain and snowmelt water are routed between overland flow and infiltrated on the basis of the component's characteristics of the soil type for the pervious zones. GAWSER employs two methods for routing the combined overland flow from impervious and pervious areas. They are an area/time versus time method (Viessman et al. 1977) or a single linear reservoir plus lag and route channel routing. GAWSER uses a linear reservoir approach to compute the outflow from subsurface and groundwater storage that the infiltrated water on impervious zones produces. The routed

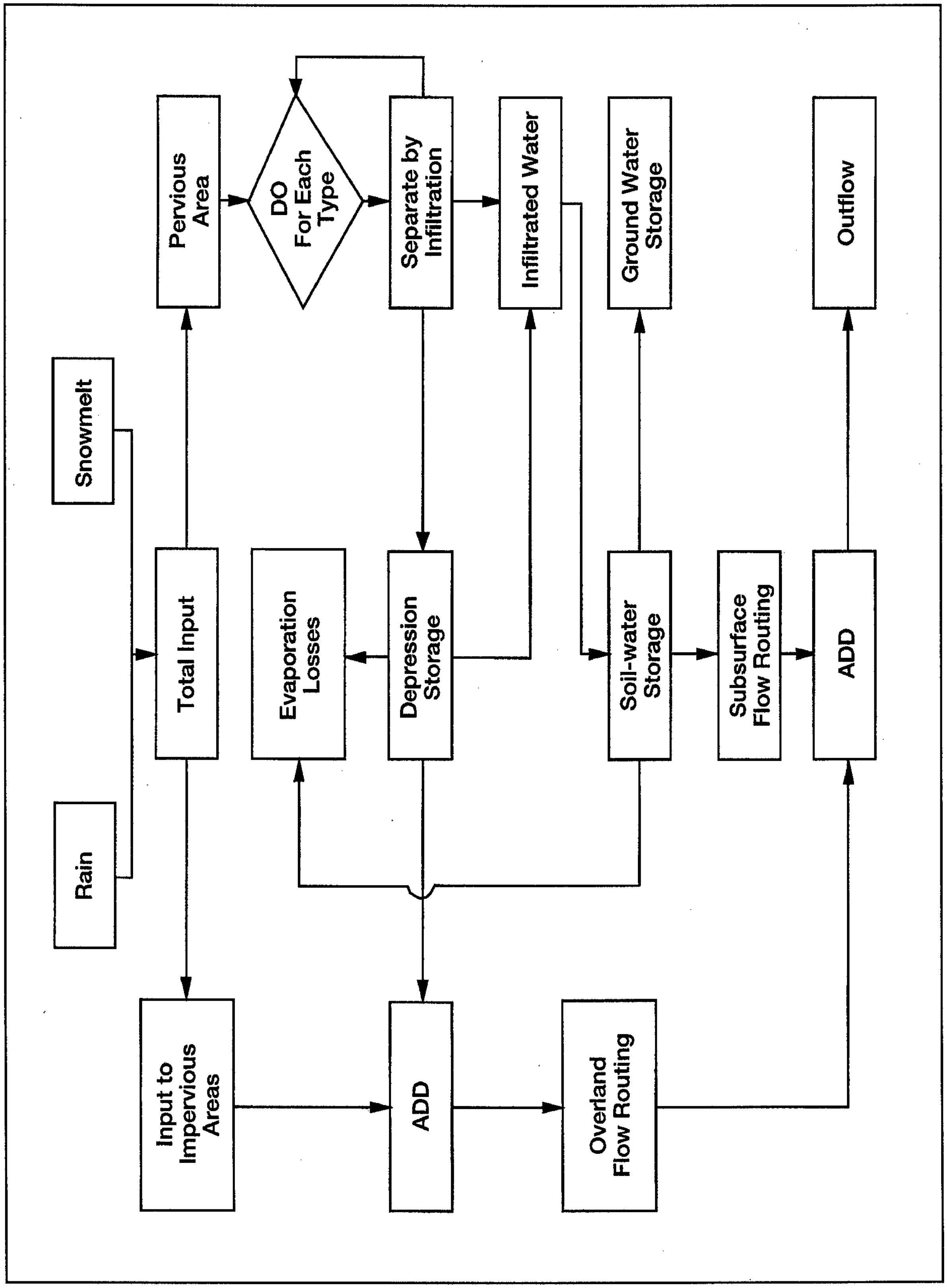


Figure F-6. Flowchart of GAWSER Subwatershed Model (after Schroeter and Whiteley 1987a,b)

overland flow, subsurface flow, and groundwater outflow are summed to produce basin discharge.

(2) The snowmelt submodel of GAWSER (see Figure F-7) is based on a simple temperature index model developed by Schroeter and Whiteley (1987a,b) and Schroeter (1988). This submodel, called the Areal Snow Accumulation-Ablation Model (ASAAM), accounts for the processes of refreeze, compaction, new snow deposition, rain deposition, snowmelt, and release of liquid water. ASAAM has also been used to simulate erosion and redistribution of shallow ephemeral snowpacks (Schroeter 1988), which has applicability in midwestern United States winter environments. Refreeze and snowmelt are calculated using a temperature index approach that employs a seasonally variable melt factor. The snowpack water content is accounted for, and all liquid water in excess of the capillary holding capacity is made available for runoff. New snow deposition and snowpack compaction are modeled by accounting for the density of new-fallen snow and the compaction effect of a settling snowpack.

(3) The snow accumulation and melt are distributed by desegregating the watershed into subwatersheds, as described previously, as well as further subdivision by Zones of Uniform Meteorology (ZUM). Therefore, each subwatershed is analyzed, and discharge is computed for each ZUM before summing to the subwatershed scale. In the case of analysis of snowmelt runoff in shallow ephemeral snowpacks, the ZUMs are further separated into blocks or elements of characteristic physical parameter types that control snowpack distribution. Examples of these block types are plowed fields, road ditches, and coniferous forests.

(4) The GAWSER program is written in FORTRAN and can be run on an IBM PC or equivalent running under MS DOS. GAWSER has been recently integrated into a system for real-time reservoir control referred to as GRIFFS (GRCA 1989).

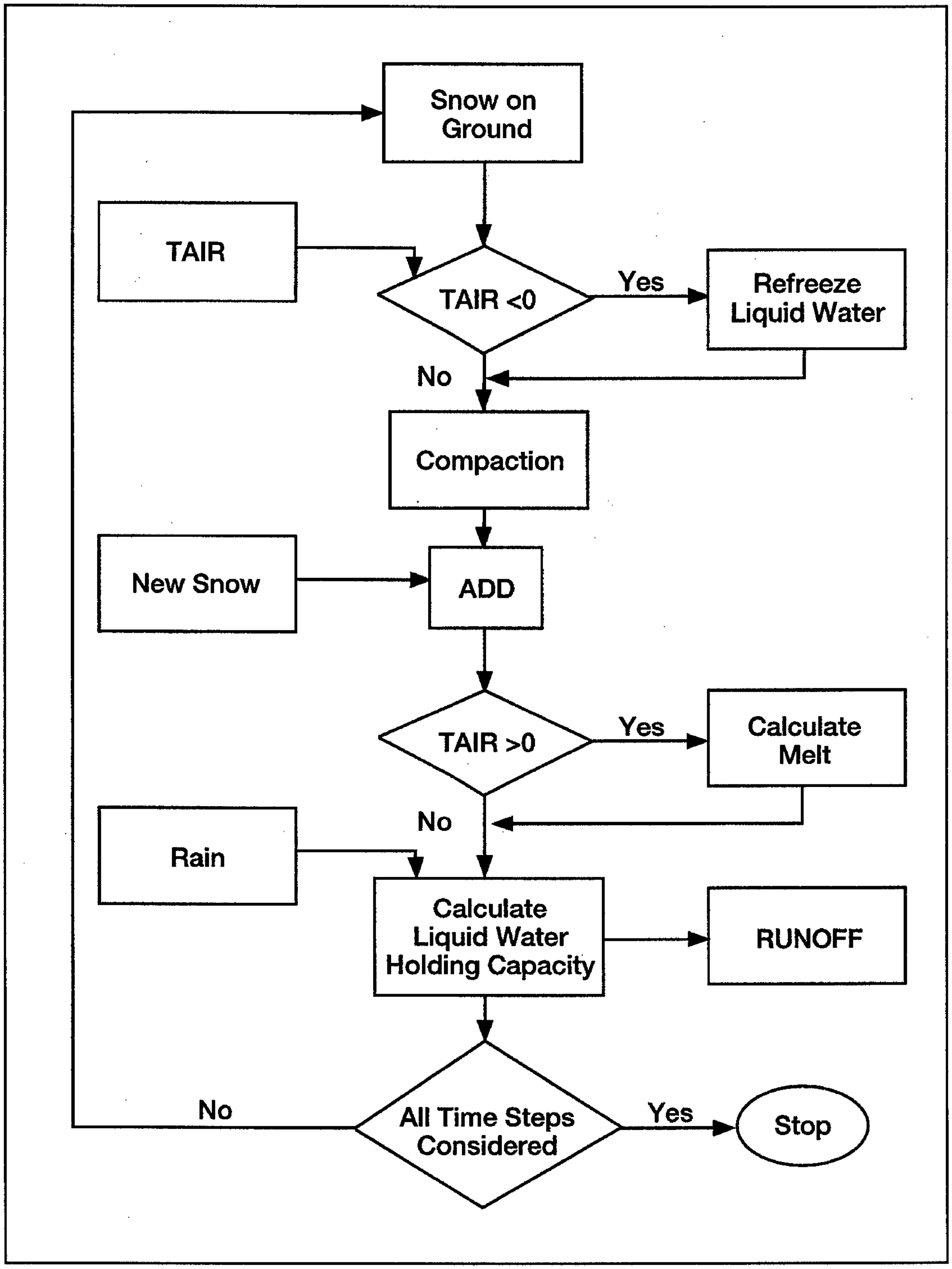


Figure F-7. Flowchart of GAWSER Snowmelt Model (after Schroeter and Whiteley 1987a,b)